Bulletin
of
The Geological Society of America

VOLUME 68 April 1957 NUMBER 4

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Subscription $15.00 per year.
Printing Office: Mt. Royal & Guilford Aves., Baltimore 2, Md.
Communications for publications should be addressed to The Geological Society of America, Dr. H. R. Aldrich, Secretary, 419 West 117 Street, New York, N. Y.
Please give notice of change of address 4 weeks in advance. Claims for nonreceipt of the Bulletin of a given month must be sent to the Secretary of the Society before the end of the second succeeding month.

Entered as second-class matter at the Post Office at Baltimore, Md., under the Act of Congress of July, 16, 1894.
Accepted for mailing at special rate of postage provided for in Section 1103 Act of October 3, 1917, authorized on July 8, 1918.

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 68, PP. 395–412, 10 FIGS. APRIL 1957

FAULT STRUCTURES IN THE FOOTHILLS AND EASTERN ROCKY MOUNTAINS OF SOUTHERN ALBERTA

BY G. S. HUME
(Address as Retiring President of The Geological Society of America)

ABSTRACT

In the southwest part of Alberta, Canada, the eastern Rocky Mountains have been considered to be fault blocks in which most of the faults, as interpreted, remain separate at depth. In the adjoining foothills to the east of the mountains drill holes show that many faults join at depth into sole faults which in relatively incompetent beds closely follow bedding planes and in more competent beds diagonally cut across the bedding. This is step faulting as described by Rich in the southern Appalachians in which the faults are directly related to the beds that are intersected. Recent drilling in the Livingstone Range of Southern Alberta has shown that this mountain is composed of superimposed folded fault slices. It is believed that the folding resulted from the faulting and is not necessarily subsequent to the faulting as has been generally assumed. The eastern Rocky Mountains, as typified by the Livingstone Range, may be composed of superimposed folded fault slices rather than individual fault blocks as has been formerly thought.

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Introduction

Statement of Problem

The character of the great overthrust faults on the eastern edge of the Rocky Mountains in Canada was first recognized in 1885 by McConnell (1887, p. 33D) in the Bow River valley west of Calgary and later in Southern Alberta as a result of Bailey Willis' (1902, p. 305) report on the Lewis Thrust in Montana. In 1954 Clark (1954b, p. 5) further described the fault at the front of the mountains in Bow River valley and named it the McConnell fault. He drew particular attention to the low angle dip of the fault on its eastern margin under Yamaska Mountain (Fig. 1) and, from studies in Bow River valley, concluded that west of the low-angle portion the fault probably dipped westward at about 45°. On this basis, since Cambrian strata are overthrust onto Upper Cretaceous, making the stratigraphic displacement about 13,000 feet, it can be calculated that the beds above the fault were originally about 4 miles west of their present location. However, it is readily apparent that the estimated amount of displacement is entirely dependent on the assumed inclination of the fault. As shown by McConnell's section (1887) along the South Fork of Ghost River, the McConnell fault is folded, a feature which is regarded as of considerable significance.

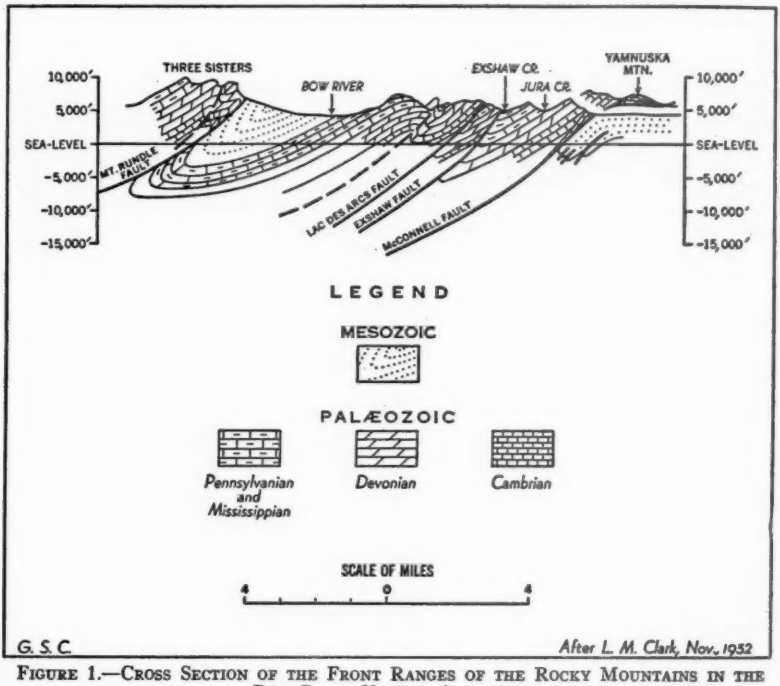

FIGURE 1.—Cross Section of the Front Ranges of the Rocky Mountains in the Bow River Valley, Alberta

The McConnell fault thus has a large displacement which is by no means unusual compared with the amount of displacement of other overthrust faults in the eastern mountains. A few faults, for example the Lewis thrust, greatly exceed it in total displacement; but most of the faults in the foothills are smaller. As presently interpreted there are differences between the McConnell fault and many of the faults in the foothills. These differences undoubtedly are related to the character of the strata that have been deformed. Indeed, the competency of the beds may be the main reason

for the relatively large size of the fault blocks of predominantly Paleozoic limestones in the eastern mountains and for the reasonably wide spacing of the faults. In the foothills, however, where the superficial strata are composed of predominantly Mesozoic shales and sandstones, the individual faults, as observed at the surface, generally have less displacement and are more closely spaced. A further apparent difference is that the Paleozoic fault blocks in the mountains are overthrust onto one another by faults that generally are not interpreted as intersecting at depth (North and Henderson, 1954, cross section); whereas in the foothills, as known from records of wells drilled for oil and gas many faults which are steeply inclined at the surface originate from sole faults, some of which are only moderately deep. The question arises, therefore, as to whether these differences are real or assumed. They may arise from incorrect interpretations based on insufficient information in the mountain areas, because the nature of the faults in the foothills now seems reasonably well established by drilling. The answer to this problem is of great economic significance.

It might be helpful, therefore, to examine a number of foothills areas where drilling has been done to attempt to unravel the sequence of events and arrive at some tentative conclusions.

Acknowledgments

The writer acknowledges with gratitude the help and suggestions received from R. J. W. Douglas and D. K. Norris of The Geological Survey of Canada and from J. C. Scott of The Husky Oil and Refining Company who are participating in the development of the Savanna Creek anticline.

Turner Valley Structure

Location and Size of Field

More than 500 wells have been drilled in Turner Valley oil and gas field, (Gallup, 1951, p. 797–821) a complicated structure on the eastern edge of the foothills, 35 miles southwest of Calgary. The Turner Valley field is about 20 miles long and 2–3 miles wide. The top of the Mississippian Rundle group, containing the producing oil and gas beds of porous dolomites, occurs at a minimum depth of 3450 feet on the crest of the structure and at a depth of more than 9000 feet on the west flank where the dip is approximately 20°–25°. Within the productive zone the difference in elevation between the highest well drilled on the crest and some drilled on the west flank is approximately 5000 feet. This is believed to be the largest productive closure within a single producing zone of any field in North America. West of the productive zone the Rundle beds continue their westerly dip until interrupted by the Outwest fault. The relief of the Turner Valley limestone block of Mississippian age is believed to be more than 7000 feet. Thus, if the Mississippian beds of the Turner Valley field were denuded of their overlying sediments, they would form a mountain in configuration closely resembling Mount Rundle at Banff from which the name of the productive formation is derived.

Step Faulting

On its eastern margin the Turner Valley field, particularly in the central section, appears to be an asymmetric drag fold, but elsewhere on the eastern margin the drag folding is not so pronounced. A well drilled on the west flank in the expectation that Devonian beds would occur under the Mississippian intersected the fault and instead of encountering Devonian beds underlying the productive structure, drilled into Lower Cretaceous strata (Fig. 2). The position of this fault for a considerable distance conforms rather closely to the bedding of the shaly Lower Mississippian Banff Formation, and then sharply cuts upward through the more massive Mississippian beds at the eastern edge of the drag fold. The bedding-plane character of some faults was noted by Beach (1943, p. 51) in the Moose Mountain area, west of Calgary. Douglas (1950, p. 90) has further described similar faults, illustrated by the major fault under the Turner Valley structure, where in incompetent beds between massive beds the fault tends to parallel the bedding planes of the incompetent members, and invariably cuts at a sharper angle across the more competent members. This gives a steplike effect to the faulting, as outlined by Rich (1934, p. 1589) in the southern Appalachians. For this reason the dips of many faults have gentle and

steeper parts according to this competency of the beds that they cut. There seems to be little doubt that the bedding-plane character of the fault under Turner Valley may be ascribed to the shaly nature of some of the Banff strata in contrast to the much more massive limestones and dolomites of the overlying Rundle Formation.

after the accumulation of gas and oil, the Turner Valley fold was faulted, and subsequently, but probably in the same episode of deformation, the Outwest fault occurred, and the Highwood uplift overrode the west flank of Turner Valley. There now seems reasonable doubt that this is the right sequence of faulting. As will be deduced from other areas, the Outwest fault may

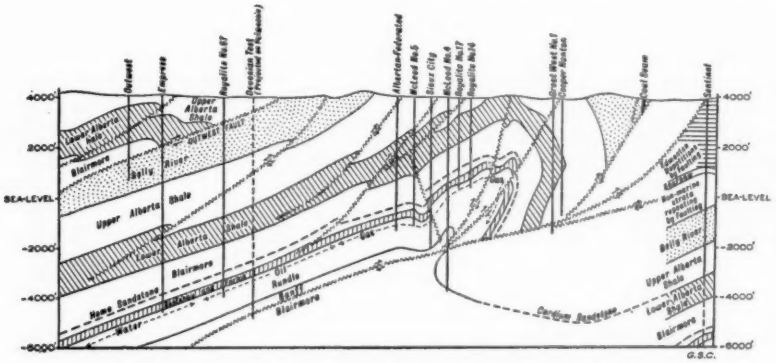

FIGURE 2.—CROSS SECTION OF CENTRAL TURNER VALLEY, ALBERTA

Outwest Fault and Highwood Uplift

When wet gas was discovered in the Mississippian beds of Turner Valley in 1924 it was logical to test other similar structures in the foothills. Thus the Highwood uplift separated from Turner Valley by the Outwest fault (Fig. 3) was drilled. It was predicted and proven by drilling that the top of the Mississippian beds in the Highwood uplift was at a higher altitude than the top of the same beds in the Turner Valley structure. However, the Highwood uplift yielded only shows of oil, gas and salt water. This led to the conclusion (Hume, 1944, p. 42; Link, 1953, p. 132; Gallup, 1951, p. 817) that oil and gas in Turner Valley had accumulated in an ancestral fold prior to faulting when the Mississippian limestone constituting the present Highwood uplift was continuous with the same limestone of Turner Valley and was located down the dip far to the west. Drilling in Turner Valley confirmed the belief that accumulation had preceded faulting, because at the north end of the field the oil-water line in one fault block is much lower than in another, although presumably at the time of accumulation they were at the same level. The inference is that, have occurred first and the Turner fault block may actually be an "under-fault" block which was shoved outward and upward from under the Highwood uplift and which carried the Highwood uplift forward on its back. This sequence implies that the order of faulting was from west to east, in conformity with the order of regional development of the mountain provinces in which the Coast Range deformation preceded the development of the Interior plateau mountain ranges and these in turn preceded the Rocky Mountain uplift.

Relation of Faulting to Gas and Oil Accumulation

The fact that the oil accumulation preceded all faulting seems firmly established. It is not the purpose here to discuss in full the timing of the original deformation which caused the ancestral Turner Valley fold in which the gas and oil accumulated. There is much evidence in support of orogenic movements at the end of the Paleozoic, whereas the faulting is definitely Laramide. It has been shown (Michener, 1924, p. 59) that the Sweet Grass arch of Southern Alberta and Montana probably originated at the end of the Paleozoic. Abundant data on the Peace River area of Alberta and British

Columbia show pre-Mesozoic deformation and faulting on a considerable scale. In the Mackenzie River region the unconformable relations between the Paleozoic and Mesozoic strata were noted many years ago (Hume, 1924, p. 10-11). Thus late Paleozoic deformation in the eastern Rocky Mountains and adjoining Plains seems firmly established. Whether the oil accumulated immediately after Paleozoic deformation or in the early stages of the Laramide revolution cannot be determined from the available structural evidence. All that can be deduced is that it preceded the main episode of Laramide faulting.

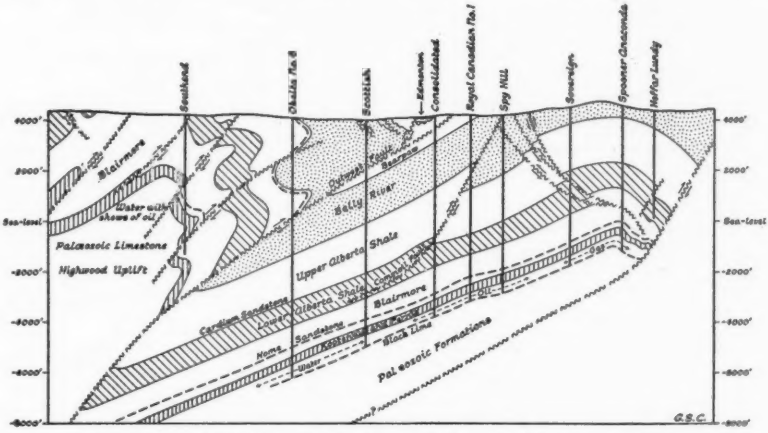

FIGURE 3.—CROSS SECTION OF SOUTHERN TURNER VALLEY, ALBERTA

JUMPINGPOUND STRUCTURE

Location

Jumpingpound is a foothills structure west of Calgary (Fig. 4). When the fold outlined by surface mapping was drilled, a sole fault, now called the Jumpingpound fault, was intersected by two wells about a mile apart, one west of the other. This part of the fold above the Jumpingpound fault contains no Mississippian strata, and hence no gas or oil was found. As a result of 1939 field studies, (Hume, 1940, p. 9, 17) predictions were made prior to further drilling that Mississippian strata occur above the Jumpingpound sole fault about 4 miles west of the crest of the superficial Jumpingpound fold and that under the Jumpingpound fault east of the crest of this fold is another fold having favorable gas and oil prospects. Subsequent seismic work followed by drilling proved both of these predictions correct; and although it may be a coincidence that the top of the Mississippian was intersected in a well under the Jumpingpound fault at a depth almost identical with that predicted, the Mississippian core of the fold above the Jumpingpound fault to the west was much deeper than anticipated (Fig. 5). (For a comparison of cross sections showing predictions and results *see* Link, 1954, p. 362–363.)

West Jumpingpound Faulted Fold

This western Jumpingpound fold containing Mississippian strata above the Jumpingpound fault is the northwestward continuation of the north end of Turner Valley, about 21 miles to the southeast. At its north end the Turner Valley fault block plunges downward to the northwest but in the west Jumpingpound area it rises along the strike. This was the basis for the prediction that Mississippian beds lie above the Jumpingpound fault. In 1943 a well in this area reached the top of the Paleozoic at a depth of 11,585 feet and in these strata there were significant shows of oil and gas. On test, however, the well yielded only salt water. From this information, therefore, it is inferred that a cross section of the Jumpingpound area should depict the steplike change from a near-bedding-plane fault to a steeper diagonal fault cutting across the Paleozoic limestone. This change is

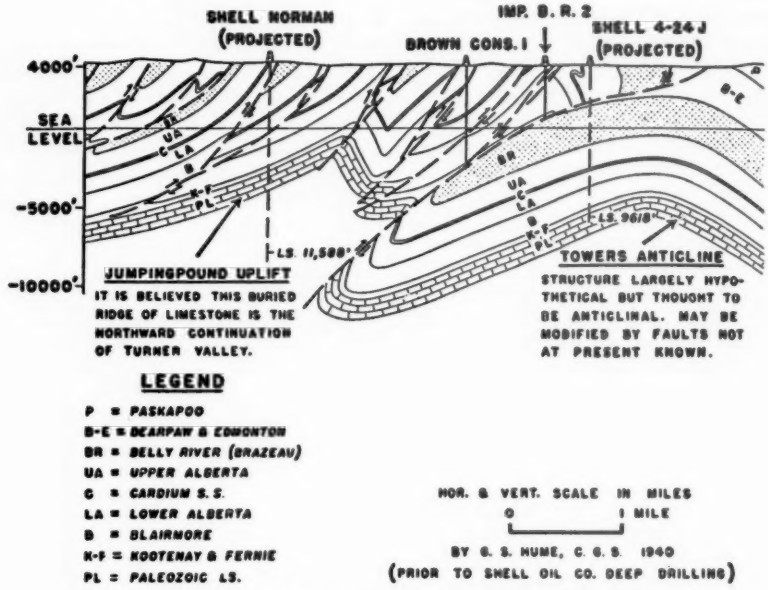

FIGURE 4.—CROSS SECTION THROUGH JUMPINGPOUND STRUCTURE

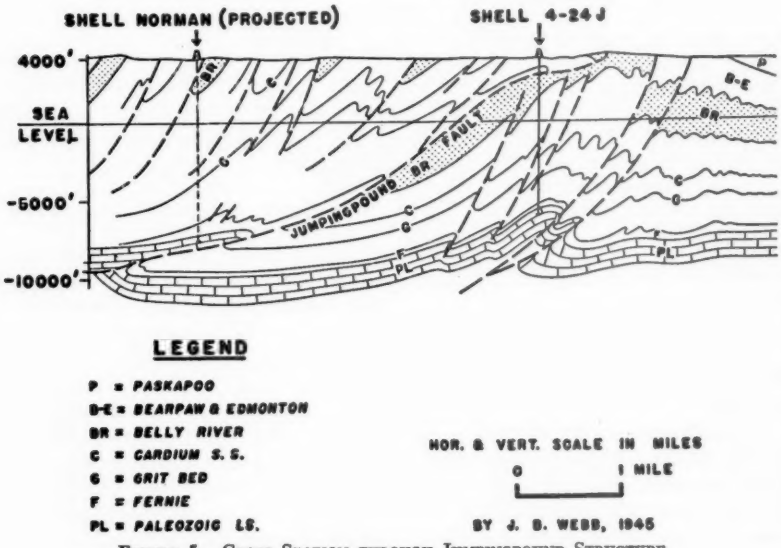

FIGURE 5.—CROSS SECTION THROUGH JUMPINGPOUND STRUCTURE

not apparent in the cross section by Webb (Fig. 5). The two wells, one on the crest of the superficial drag fold and the other about a mile west of it, drilled into Fernie shale of Jurassic age above the Jumpingpound fault and hence the fault east of the Paleozoic limestone in the west Jumpingpound fold closely follows the bedding of those strata for about 4 miles. West of this part in which the dip of the fault plane is gentle the fault is presumably steeper where

it cuts the Paleozoic competent beds. The part of the fault near the surface is also steeper than the near-bedding-plane part.

Economic Significance of East Jumpingpound Faulted Fold

The economic importance of the fold on the eastern edge of the east Jumpingpound area under the Jumpingpound fault need not be stressed. The productive field now extends for more than 10 miles. It is a gas field producing distillate and the gas contains significant amounts of hydrogen sulfide from which sulfur is being extracted. The reserves of the field are calculated to be 500 billion cubic feet. Production to July 1, 1956, was 59 billion cubic feet containing 525,000 barrels of distillate and 209,000 barrels of gasoline; 55,000 short tons of sulfur were recovered.

Lewis Fault

Location and Displacement

The Lewis fault of Montana and southwestern Alberta was first described by Willis (1902, p. 305–52). It has been studied by Campbell (1914, p. 12) in Montana, by Daly (1912, pt. 1, p. 90) at the International Boundary, and has been the subject of discussion by many authors in Canada including Mackenzie (1922, p. 110) and Clark (1954a, p. 105). The Clark Range in Montana and its northward continuation in Canada, an area about 100 miles long and 20 miles wide consisting of Proterozoic strata has been overthrust onto Mesozoic and Paleozoic strata. Campbell considered the displacement in the area of the southern end of Glacier National Park to be about 15 miles, whereas Daly considered the displacement at the International Boundary to be as much as 40 miles. In Canada there are strong seepages of oil through the Beltian rocks on their east edge on Cameron Brook near Waterton Lake and on their west edge in a number of valleys in streams flowing westward into Flathead River. A well drilled in 1901 on the east side intersected Mesozoic shales under Beltian rocks, and a well completed in 1953 in the Flathead area penetrated 4500 feet of Beltian strata and intersected Mississippian strata below the fault. Gas and oil were present in small amounts in the Mississippian, but porosity necessary to provide reservoir conditions was lacking.

Syncline under Clark Range

The position of the Lewis thrust under the Clark Range can thus be postulated with a fair degree of assurance although there still may be differences of opinion as to the interpretation of complicated relationships in which the Lewis thrust is apparently cut by another fault along the east side of Flathead valley as shown in Clark's cross section (Fig. 6). There is little reason, therefore, to doubt that the part of the Lewis thrust under the Clark Range conforms to a considerable degree to a bedding-plane fault and that it is regionally synclinal.

Time Relations to Foothills Faults

From interpretations made, since foothills structures can be traced under its eastern edge, the general assumption has been that the Clark Range is younger than the foothills faulting and hence overrode the foothill structures. This conception presents some real difficulties because, besides involving different periods of deformation or at least different stages in the same period of deformation, it involves bevelling off the foothills and thrusting the Lewis fault slice onto an erosion surface as postulated by Willis (1902, p. 336). A further problem which need not be discussed at this time, but which should not be ignored in any final interpretation, is the normal faulting on the east side of Flathead valley, which obviously cuts and displaces the Lewis thrust. Some of the complications of structure may be due not to the Lewis thrust overriding foothills structures but to the fact that these structures developed below the Lewis thrust in a fault slice formed subsequent to it and moved upward and outward (northeastward) from under it.

East of the outcrop of the Lewis thrust a few miles north of Waterton is the Pincher Creek wet gas field. Gas has been found in Mississippian limestone at a depth of 11,000–12,000 feet. The field is important as it is estimated to contain a reserve of more than 3 trillion cubic feet of natural gas and large amounts of other

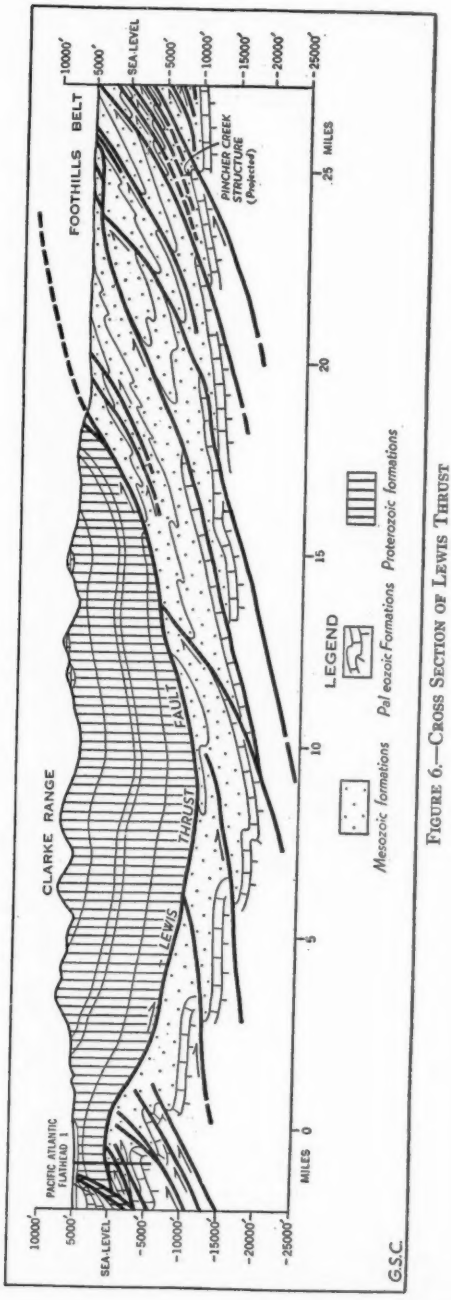

FIGURE 6.—CROSS SECTION OF LEWIS THRUST

After L. M. Clark

hydrocarbons. The question at once arises as to whether this is a fold like Turner Valley in which the accumulation preceded the faulting and whether the next projected fault block to the west of it will be, like the Highwood uplift west of Turner Valley, filled with water only. This economic application of the sequence of geological events is thus of very great importance, but present data are insufficient to provide a positive answer. However, the Turner Valley geology which seems reasonably well understood, may be worthy of consideration as a significant guide.

Savanna Creek Area

Livingstone and Highwood Mountain Ranges

A mountain ridge, known as the Livingstone Range, extends north for many miles from the Crowsnest Pass area of southwestern Alberta east of Blairmore (Fig. 7). The range, which in the south consists of a single ridge, is capped by Mississippian limestones. East of it is the foothills belt, 8–10 miles wide. West of it, also about 8 miles wide, is a series of ridges composed of Mesozoic rocks of which Crowsnest Mountain is a prominent erosional remnant of the overthrust which underlies Highrock Range and is the northwest extension of the Lewis thrust. Highrock Range in the Crowsnest Pass area is the continental divide of the Rocky Mountains between Alberta and British Columbia. The ridges of Mesozoic strata east of Highrock Range and west of the Livingstone Range are apparently no different than the ridges of Mesozoic strata of the foothills to the east. Thus the Livingstone Range has been considered to be a mountain inlier within the foothills. It undoubtedly owes its altitude, at the maximum above 8000 feet, to the fact that its Paleozoic limestone capping has been more resistant to erosion than the Mesozoic beds in the ridges both east and west of it. Thus it would be expected that it would be no different structurally from the foothills and eastern mountains.

In the south, the Livingstone Range has an overthrust fault on its eastern margin, but farther north it is anticlinal. About 40 miles north of Crowsnest Pass near Sentinel Mountain and Hailstone Butte, it again is faulted on its eastern edge by the Hailstone Butte fault. West of this area there is a syncline with an anticline still farther west in the Savanna Creek area of the Highwood Range (Fig. 7). This fold in the area north of Savanna Creek has been drilled, and two wells have found potential gas production in substantial amounts. Another well is drilling.

Structure as Revealed by Drilling

Drilling in this area was begun in the expectation that below the Mississippian cap rock the Devonian could be tested. However, in the first well drilled on Dry Creek at an altitude of 6330 feet, the Livingstone fault was intersected at a depth of 140 feet (Fig. 8) and the well passed into Upper Cretaceous strata (Irwin, 1955a, p. 18; Irwin, 1955b, p. 173). A subsequent well drilled near the first one cut the Mesozoic beds underlying the Livingstone fault, reached the Mississippian beds again at a depth of 4400 feet, and in a porous zone obtained an uneconomic flow of gas (Scott, 1953, p. 138). Drilling was continued, another smaller fault was intersected at a depth of 5235 feet, and the well passed back into the lower part of the Lower Cretaceous. The top of the Mississippian limestone was again reached at a depth of 6400 feet, and in these beds between depths of 7210 and 8150 feet flows of gas aggregating 50 million cubic feet a day were obtained.

A third well was drilled a mile east of the others. The Livingstone fault, as interpreted, was intersected at a depth of 1454 feet in contrast to 140 feet in the first well; conclusively proving that the fault is folded. A second fault which is regarded as being the deeper fault drilled at a depth of 5235 feet in the productive well was also intersected at a depth of about 4450 feet. The well was completed as a substantial potential producer at a depth of 8481 feet.

Opinion differs concerning the second fault intersected in each of the two wells at 5235 and 4450 feet, respectively. R. J. W. Douglas (Personal communication) regards this as a fault within the Dyson Mountain thrust sheet which directly underlies the Livingstone thrust sheet, whereas Scott (1953, p. 138) and others think it is the Dyson Mountain fault under-

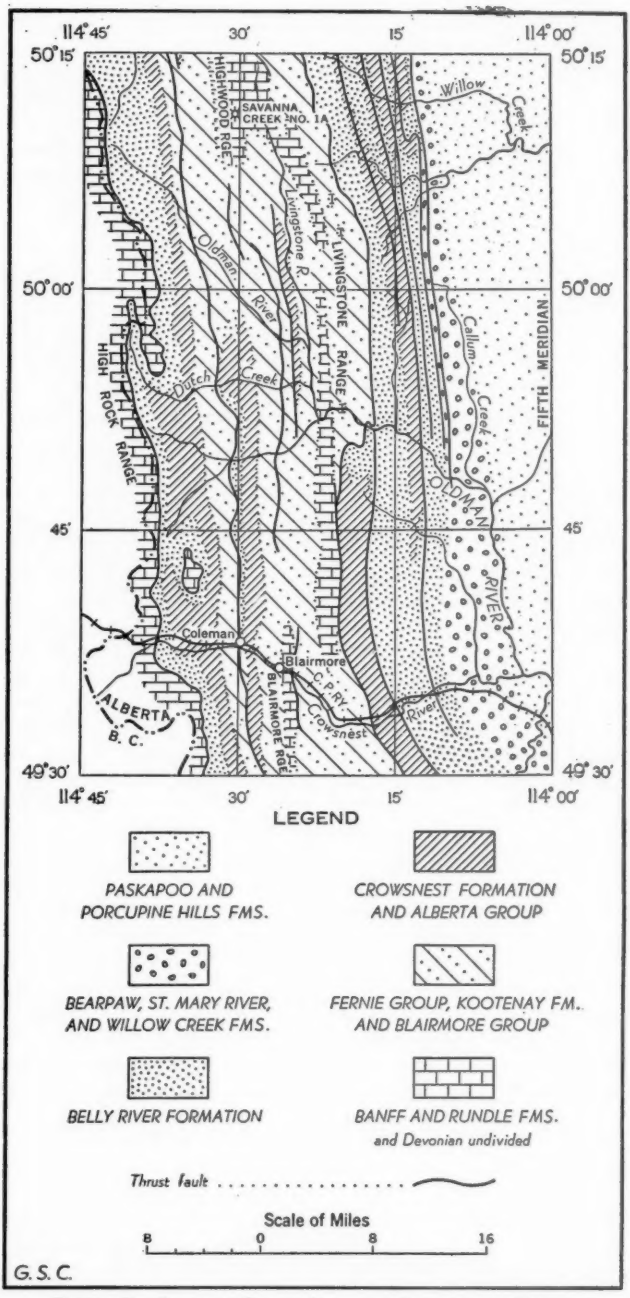

FIGURE 7.—SAVANNA CREEK AREA AND LIVINGSTONE RANGE

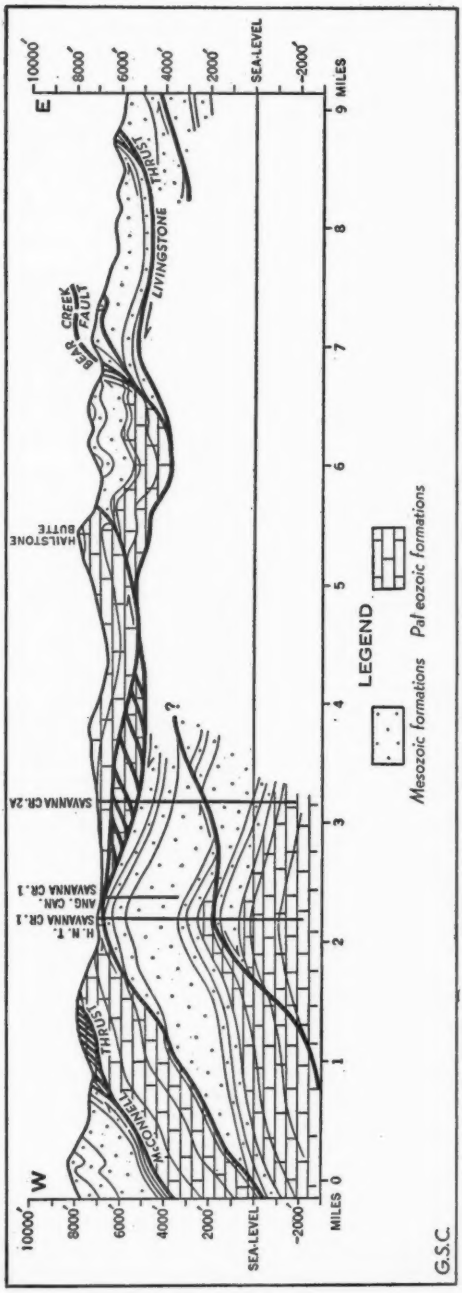

FIGURE 8.—Cross Section of Savanna Creek Structure
After D. K. Norris

lying the fault sheet of the same name. Obviously it was regarded by Irwin (1955b, p. 20) as the Dyson Mountain fault, and he states,

> "In the highest, largely exposed fault plate more than 100 feet of Rocky Mountain formation (Permo-Pennsylvanian?) is present and undoubtedly some has been removed by erosion, as it forms a mountain top. In the next underlying fault plate beneath a depositional contact only 60 feet of Rocky Mountain formation is present. In the third occurrence of the Rundle Jurassic strata directly overlie Mississippian strata so that no Rocky Mountain formation intervenes, and the more familiar disconformable and erosional contact between the Mesozoic and Paleozoic systems, as known in the foothills and plains, is present. These relationships give evidence, and some measure of the progressively greater amount of eastward movement in each successively higher and younger fault plate".

This seems to indicate that the second fault is the Dyson Mountain fault rather than a minor fault within the Dyson Mountain fault slice, since on a minor fault the thickness of the Rocky Mountain formation would be expected to be the same on both the upthrown and downthrown sides. However, such a conclusion is dependent on a distinction between Rocky Mountain and Rundle strata in a succession where such a division is not easily recognized from well samples. Thus there is still reasonable doubt that the Dyson Mountain fault has been intersected by the well.

It is generally agreed, however, that there are two fault slices, the Livingstone and the Dyson Mountain, in the fold. The trace of emergence of the faults in the foothills east of the mountains has been mapped. The Hailstone Butte fault, on the east side of the Livingstone Range, thrusts Paleozoic limestones onto Jurassic shales. It is either a branch of the Livingstone fault or the projection of the fault cut in Well No. 1A at a depth of 5235 feet and in Well No. 2A at a depth of 4430 feet. In this case it would be interpreted as a branch fault from the Dyson Mountain sole fault and would cut and offset the Livingstone fault which outcrops about 6 miles east of the wells and thrusts basal Lower Cretaceous strata onto Upper Cretaceous (Belly River) rocks. The interpretation of these detailed features will be clarified by the results of drilling now proceeding or contemplated at various sites on Plateau Mountain which has an elevation of more than 8000 feet.

Superimposed and Folded Fault Slices

In spite of incomplete understanding of this structure, there is sufficient information from which to draw some tentative conclusions. The significant feature is that this mountain mass is composed of folded fault slices superimposed on one another. There are many reasons to suppose that deeper fault slices may also be present. This conception of mountain structure is much different from that formerly held for the Livingstone Range, but it is in conformity with knowledge of foothils structures as determined by drilling. The Livingstone Range is considered to be an inlier of the foothills. Perhaps this fact might leave some doubt that its structure is in general different from the mountains farther west.

Eastern Rocky Mountains of Similar Type

However, northwestward from the Savanna Creek area the Highwood Range becomes structurally indistinguishable from other ranges in the eastern Rocky Mountains, and hence the conclusion that all are essentially similar in structure is inescapable. Differences in detail according to the competency of the beds in each particular area are to be expected. The pattern of the McConnell fault in the Bow River valley fits such a conception of structure because of its flat near-bedding-plane displacement in some parts and steeper diagonal faulting across the beds in other parts, as well as folding of the fault as shown in a cross section published with McConnell's 1886 report.

Folded Faults and Sequence of Fault Slice Formation

Fault-Slice Movement under Heavy Load

Although the fault slices of the Livingstone Range involve some massive and hence reasonably competent Paleozoic limestone beds, they are largely composed of less competent sandstones and shales of Mesozoic age which vary widely in competency. Coal seams occur in the sequence containing the most massive sandstones. Coal seams or coal-bearing zones in the faulted foothills belt are commonly the loci of slipping planes, and an Upper Cretaceous section composed largely of shales is readily con-

torted under deformative stresses producing faults. The marvel is that these fault slices have been relatively slightly deformed even though they have been displaced many miles on near-bedding-plane faults. Minor distortions and thickening of beds, however, are common in the less competent members. The most satisfactory explanation is that when they were moving forward and upward they were under such a load that each acted as a rigid unit.

Sequence of Fault-Slice Formation

It has generally been assumed that the strata faulted under severe compressive stresses and that to the west each fault block in succession overrode the one east of it. In this way a number of fault blocks were formed in which the faults remained separate at depth. Such a conception of origin is not tenable on the basis of comparatively high rigidity in the fault slices produced by a superimposed load. Under these conditions it seems much more probable that each fault slice emerged to the east from under the preceding overlying mass and carried it forward by slipping on a deeper fault plane. Faulting within such a fault slice also occurred by the development of "back limb" thrusts originating from the deeper underlying fault. The movement on such a sole fault may partly explain why the fault planes in the foothills are to a great extent confined to a thin zone. In fact, the sharp definition of the faults commonly each as a single plane is regarded as highly significant. It also may partially explain the folding in some faulted foothills areas where the faults as well as the strata involved with them are highly folded.

Folded Fault Areas

One notable folded fault area in the foothills is the Brazeau structure, southwest of Edmonton, in which the folded nature of the faults was originally described by Hake, et al. (1942, p. 291–334) and later by Mackay (1941, p. 189–190) and Scott (1951, p. 2325–2347). However, folded faults are elsewhere present and have been described by a number of authors. Deiss, (1943, p. 1158–1159) particularly referred to them in the Sawtooth Range, Montana. All this folding is not thought to be due to a single cause. Some of it, particularly that of a regional nature, may well be folding subsequent to faulting, but in the foothills in Canada much of it is believed to be directly related to the faulting itself. Two factors are involved: the relationship of the faulting to near-bedding-plane thrusts as has been so well described by Douglas (1950, p. 84) and the effect produced by faulting or crumpling in lower fault slices as they move forward under the load of an overlying mass.

Relationship of Folding to Faulting

As pointed out by Douglas (1950, p. 84) "folding of these thrust planes has taken place where considerable bedding-plane thrusting is involved or especially where bedding and thrust planes are essentially parallel". Reference has already been made to the steplike character of these faults. It is inevitable therefore, as shown by Douglas, that at the point where the dip of the fault changes upward from a steeper to a gentler slope (Fig. 9), an anticline will develop and produce not only a fold in the fault itself but in the beds above and below it. In such cases the folding of the fault and adjoining beds is contemporary with the faulting, not subsequent to it as has been so frequently assumed. In no way is this a recumbent fold resulting in a fault under great deformation, but it is a fold resulting from a change in inclination of the fault plane and hence is directly related to the character of the bedding traversed by the fault. It is believed that the Savanna Creek fold is this type. The west Jumpingpound Paleozoic structure extending northwestward is another example.

It might logically be assumed that, since an upwarp or anticline develops where the dip of the fault changes upward from steep to less steep, a syncline would develop at the point where the dip changes upward from a less steep to a steeper part. This is, however, not believed to be the case. This change upward from a less steep to a steeper part where the fault cuts across more competent beds produces a drag fold in the competent strata and this is reflected in the overlying strata as well. This is because the beds below the steeper part of the fault offer additional resistance because of their

competence, to forward and outward movement. Such drag folds as deduced from field evidence can occur in Mesozoic as well as Paleozoic strata according to conditions prevailing where there are marked contrasts in competency between adjoining zones of strata. The superficial fold in Mesozoic strata above the Jumpingpound fault at Jumpingpound is an example of this type of drag-fold structure, and Turner Valley with its drag fold on the eastern edge in Paleozoic limestone is another.

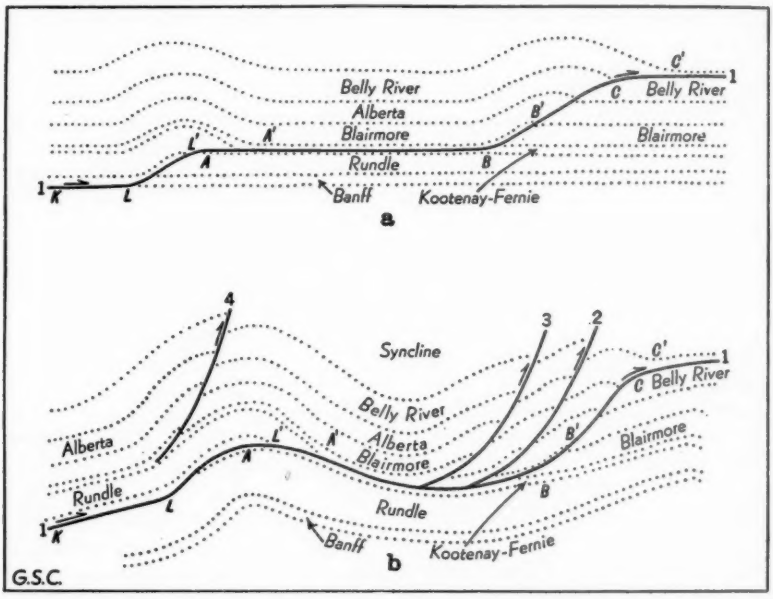

FIGURE 9.—DIAGRAM ILLUSTRATING STAGES (a, b) IN FORMATION OF FOLDED FAULTS (1–1) AND BACK-LIMB THRUST FAULTS (2, 3, 4) ABOVE INITIAL THRUST
a. Initial stage showing relation of fault habit to stratigraphy
b. Later stage showing folding of fault plane and strata above and below fault

Regional Synclines Produced by Faulting

Between the fold or anticline formed at the point where the dip of the fault decreased upward and the drag fold formed at the point where the steepness of the fault plane increases upward is an area which in general is synclinal. The syncline is formed by the east limb of the fold that develops where the dip of the fault becomes less steep upward and by the west limb of the drag fold. Since this synclinal part of the fault extends for many miles parallel or nearly parallel to the bedding planes of the strata the syncline is regional and can be easily recognized and mapped. This feature is quite apparent in certain areas of the foothills and may be illustrated by the Dyson Mountain fault under the Dyson Mountain fault slice as mapped by Hage (1946, Map 827A). In this case there is no drag fold on the eastern margin. It may have been removed by erosion or more probably it may never have been present. This may also be the explanation for the regional synclinal part of the Lewis thrust under the Clark Range, although this folding has generally been considered subsequent to faulting.

Underthrusts Indicating Changes in Fault Inclination

One feature of this drag-fold development that is very interesting from a practical point of view because it can be used in surface mapping to show the change of steepness in the sole

fault at depth is the tendency for an eastward dipping or underthrust to develop where the fault changes in inclination upward from a bedding plane to a higher angle. Such faults were shown in the cross section of southern Turner Valley (Fig. 3). These faults in this particular case can be detected only by very detailed studies of the shale cuttings of Colorado age. Because of this, their presence has been questioned. There is no doubt, however, that such faults occur elsewhere. Two were described (Hume, 1932, p. 53) in the foothills west of Calgary. The deductions to be made from such faults are economically important. If at the place where it changes its dip from diagonal across the beds to nearly parallel to the bedding planes the fault has not cut sufficiently deep in the geological section to intersect the possible oil-producing strata, it will not intersect these oil and gas beds along the bedding-plane part of the fault until it again steepens sharply to cut deeper. This distance, as has been shown for some of these bedding-plane faults, is a matter of miles. If the beds are tilted westward as are most beds in the foothills, the depth to the possibly productive beds where they are cut by the fault, and thereby thrust upward into a structure capable of containing oil and gas, may be excessive.

Back-Limb Thrust Faults and Imbrication

Occurrence on West Limb of Folds

Douglas (1950, p. 89) has discussed what he has called back-limb thrust faults in reference to his field work in the Callum Creek, Langford Creek, and Gap map areas (Fig. 9). There are, as he points out, many faults with anticlines below them instead of synclines as expected in faulted folds. Such a faulted anticline is that which is producing gas under the eastern edge of the Jumpingpound fault at Jumpingpound (Figs. 4, 5). Back-limb faults, none of which have developed from folds, have been repeatedly noted on the west limbs of anticlines in the foothills. Douglas ascribes the origin of these faults to greater differential movement between the beds on the western limb of an anticline than between those on the eastern limb. This is undoubtedly an important factor in such a break, but, if the postulated sequence of fault-slice development in which lower fault slices moved outward and upward from under higher fault slices is acceptable, there is probably another and major cause.

Growth Upward from Sole or Step Fault

Since in some instances a steplike change in a fault from a diagonal steeper part to a bedding-plane or less steep part will produce an anticline, not only in the fault itself but in the overlying and underlying beds, such a folding will tend to produce an anchoring effect and resist further movement along the fault plane, which, as indicated, is itself folded into the anticline. In such a case the stress tends to become concentrated on the west limb of the anticline, and the ultimate inevitable break develops outward from the sole fault in the lower fault slice as it was thrust forward. Such a break in the lower fault slice grows upward from the sole fault. It is resisted by the superimposed load of the overlying fault slices and although it need not, and probably in most cases will not effect a break or fault in these overlying slices, it certainly will distort them in the zone above the fault. A series of such breaks or faults in a fault slice could produce the imbrication which is so prominent a foothills feature. The sequence of development is for the sole fault to precede the imbrication.

Imbrication Related to Character of Strata

The amount of imbrication that develops is thought to be directly related to the character of the beds in the fault slice above the sole fault. If the fault slice includes Paleozoic limestones as well as Mesozoic sandstones and shales, the massive limestone beds provide a stiffening effect which results in fewer breaks than where only less competent Mesozoic beds are included in the fault slice. Thus in many structures the buried eastern edge of the Paleozoic limestone, which provides the best prospects for oil and gas in the foothills, may in a rough way be defined at depth by surface mapping by indicating areas of extremely imbricated structures in front of it in contrast to less faulted areas where its presence at depth has provided more strength. There are many areas in the foothills where this relationship is

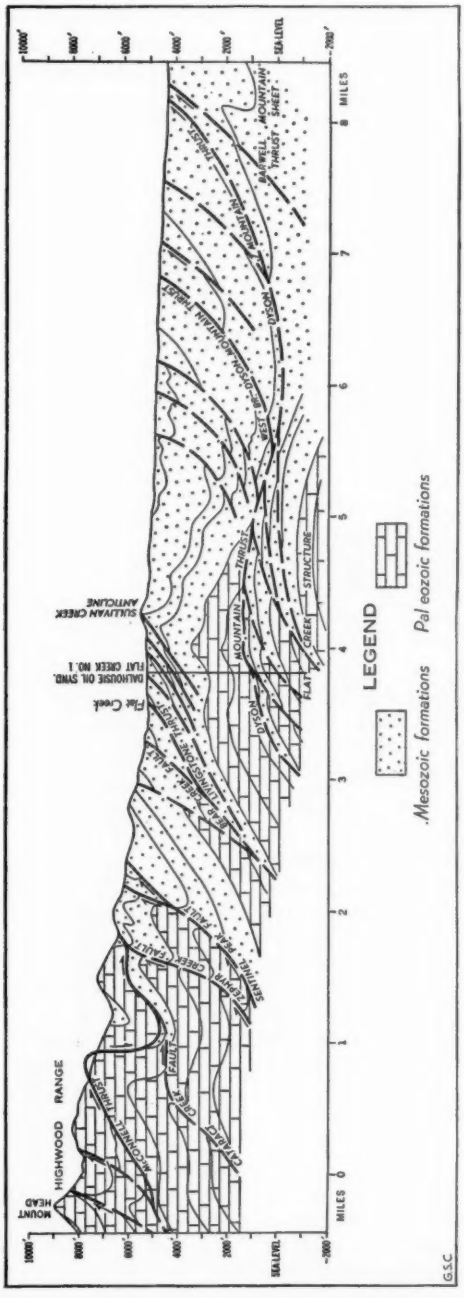

FIGURE 10.—CROSS SECTION OF SULLIVAN CREEK AREA
After R. J. W. Douglas

suspected to exist, but proof in the absence of drilling is lacking at present. It is not always possible to map all the faults in a highly imbricated shale area, but where faulting produces numerous repetitions of strata which contain some harder member in the shales, as for example the Cardium sandstone, the faults can be mapped with some degree of assurance (Hume, 1942, Map 698A).

Distortion of Overlying Faults and Fault Slices by Back-Limb Thrusts in Lower Fault Slices

The distortion produced in overlying fault slices by development of back-limb faults in lower ones deserves further study. It would be unwise at this time to ascribe the extremely folded nature of the faults in the Brazeau area to such a cause although it is suspected that it has been a major factor in their later stages of development. For example, in a cross section (Fig. 10) through the Sullivan Creek area in the foothills north of Highwood River Douglas shows a sharp fold with a window through the McConnell fault, east of Mt. Head. If the Cataract Creek fault is a back-limb thrust in a lower fault slice above the Dyson Mountain thrust, then its growth upward might be logically expected to produce distortion and folding of the McConnell fault and thus produce the asymmetrical fold as shown in the section. This implies that the fault did not originate from the fold but rather that the folding was produced by the faulting. On this basis, the interpretation of the section, particularly in reference to the Dyson Mountain thrust, might be considerably modified.

Certain features of folded faults become inexplicable without the assumption that folding was subsequent to faulting. The broadly synclinal nature of the Lewis thrust under the Clark Range as previously discussed may be of this type, but even this would be difficult to prove. Most of the folding of the faults in the foothills is believed to be directly connected with the type of faulting and is thought to be directly related to the competency of the beds.

Summary

A summary of faulting in the Foothills and Eastern Rocky Mountains may thus be stated as follows:

(1) Deformation in the Eastern Rocky Mountains and foothills produced step faulting without much apparent antecedent folding.

(2) Step faulting is the fault habit and is directly related to the stratigraphic succession of zones of relatively competent and incompetent beds.

(3) Where such a fault changes from a steeper to a gentler inclination upward that is, where it cuts less competent rather than more competent beds, the thrust-fault movement produced folding not only in the fault itself but in the strata both above and below it.

(4) Where such a fault changes from a gentler to a steeper inclination upward that is, where it cuts more competent rather than less competent beds, the thrust fault movement produced a drag fold because of the resistance to the fault movement by the more competent beds on the footwall side of the fault.

(5) Between the two folds thus produced, the structure along the near-bedding-plane part became regionally synclinal; the syncline is produced contemporaneously with and results from the fault movement.

(6) The growth of the folds in the fault movement produced an anchoring effect resisting further movement causing the formation of back-limb thrusts.

(7) These back-limb thrusts developed behind the folds of the step fault, and hence all originate from the step or sole fault and are subsequent to it.

(8) There is some evidence that the step faulting of underlying fault slices may be younger than the faulting of the overlying slices. This implies that the lower fault slices were thrust outward and upward from under higher thrust slices and that the younger fault slices are the more easterly ones. The order of development is from west to east.

(9) The development of back-limb thrusts in lower fault slices did not necessarily produce a projection of these faults into the next overlying fault slice. It, however, did produce deformation in the fault overlying the lower fault slice as well as in the beds above it in the overlying fault slice. Hence minor but relatively sharp folds, as now observed, in the major sole faults indicate back-limb faults in the next underlying fault slice.

(10) Not all folded faults should be ascribed to folds produced contemporaneously and as a result of fault movement. There may be some subsequent regional folding.

(11) Mountains like the Livingstone Range are made up of folded fault slices superimposed on one another. The folding resulted from the steplike fault movement and was contemporaneous with it.

(12) There is no apparent difference between the Livingstone Range and other ranges in the Eastern Rocky Mountains. Hence these may not be single fault blocks, as has sometimes been assumed, but like the Livingstone Range may be composed of a series of fault slices superimposed and folded over one another.

(13) The change in thickness and character of the various formations in the superimposed fault blocks gives a means of measuring the extent of displacement of the faults, the cumulative amount of which is a measure of the superficial shortening of the earth's crust due to the fault movements.

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Manuscript Received by the Secretary of the Society, November 5, 1956

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 68, PP. 413-420, 1 PL. APRIL 1957

FRENCHMAN FORMATION OF EASTERN CYPRESS HILLS, SASKATCHEWAN, CANADA

BY W. O. KUPSCH

ABSTRACT

In the eastern Cypress Hills the Frenchman Formation (Upper Cretaceous), a correlative of the Hell Creek in Montana, consists of a clay lithosome and a sand lithosome. Study of sections shows that either lithological unit may form the lowest unit of the Frenchman and that they may alternate vertically. Equivalence of the two lithologies is suggested also by their similar fossil content.

The Frenchman is distributed over a larger area than previously assumed. Certain sands formerly interpreted as belonging to the Eastend Formation (the Fox Hills of Montana) are assigned to the Frenchman because of their fossil content. Differences in grain size, color, accessory minerals, stratification, and other characteristics can be used to distinguish Eastend sands from the sand lithosome of the Frenchman.

The Frenchman rests on a surface of erosion; the upper contact is abrupt but conformable.

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INTRODUCTION AND ACKNOWLEDGMENTS

The Frenchman Formation of southwestern Saskatchewan presents three problems: (1) sands belonging to the Frenchman Formation are not everywhere easily distinguishable from sands belonging to the stratigraphically lower Eastend Formation; (2) the relationship of two different lithostratigraphical units of the Frenchman has been differently interpreted by various workers; (3) some geologists described the upper boundary of the Frenchman as conformable with the overlying Ravenscrag, others suggested a disconformable relationship.

The writer wishes to thank Mr. S. P. Jordan and Mr. N. L. McIver who assisted in measuring stratigraphic sections in the field, and Mr. B. K. Smith who conducted heavy-mineral and mechanical analyses. Mr. R. Rudichuk of the University of Saskatchewan prepared the map for publication. Thanks are also due the Saskatchewan Research Council for financial aid. The author is indebted to Dr. F. H. McLearn, Dr. L. S. Russell, and Mr. William E. Benson for offering valuable criticism of the manuscript.

GENERAL STRATIGRAPHY

Rocks of late Cretaceous and early Tertiary age are well exposed in the eastern part of the

Cypress Hills, especially in the valley of the Frenchman River (formerly called White Mud River) between Ravenscrag and Eastend, Saskatchewan. The stratigraphic section in this valley attracted geologists as early as 1883, and a lithologic description was published (McConnell, 1885, p. 27c). White sands and clays, now referred to as the Whitemud Formation, constitute the most conspicuous unit of the section. These and all overlying strata were assigned to the Laramie by McConnell, whereas he designated the underlying sands as Fox Hill. The Pierre shales are at the base of McConnell's section, the division of which has undergone many changes introduced by various workers (Table 1).

TABLE 1.—DEVELOPMENT OF STRATIGRAPHIC NOMENCLATURE IN THE CYPRESS HILLS

McConnell, 1885	McLearn (*in* Fraser *et al.*, 1935)		Furnival, 1946	Correlatives in adjacent areas of the U.S.A.
Laramie	Ravenscrag	Upper	Ravenscrag	Fort Union
		Lower	Frenchman	Hell Creek
	Whitemud	No. 4 zone	Battle	Colgate member / Fox Hills
		No. 3 zone	Whitemud	
		No. 2 zone		
		No. 1 zone		
Fox Hill	Eastend		Eastend	
Pierre	Bearpaw		Bearpaw	Bearpaw

Detailed work was done by McLearn in the Eastend area, and the results were summarized in Fraser *et al.* (1935). McLearn used the name Bearpaw Formation for the Upper Cretaceous marine shales, which were referred to as Pierre by McConnell. The overlying very fine sands and silts, representing the youngest marine deposits of the plains, were no longer called Fox Hill but Eastend. The name Whitemud was used for the conspicuous white sediments which were treated as a separate formation. McLearn divided the Whitemud Formation of the Cypress Hills into three members, or zones as he called them. The lowest consists of gray kaolinitic sand, the middle of brown carbonaceous clay and impure lignite, and the upper of refractory white, gray, and mauve clays, sandy in places. In addition McLearn recognized an uppermost or fourth member consisting of dark, bentonitic, nonrefractory shales. Lithology and stratigraphic position indicate that the Whitemud Formation is the Canadian equivalent of the Colgate member of the Fox Hills Formation in Montana and North Dakota.

Furnival (1946) adopted McLearn's divisions for the lower part of the section but removed the uppermost dark-shale member from the Whitemud and called it the Battle Formation. Through his work the Lower Ravenscrag became known as the Frenchman Formation, whereas the Upper Ravenscrag became simply Ravenscrag Formation.

DEFINITION OF THE FRENCHMAN FORMATION

Furnival (1946, p. 94) defines the Frenchman Formation as the strata lying above the Battle Formation and underlying the restricted Ravenscrag Formation. The type area of the Frenchman is along the Frenchman River between Ravenscrag and Eastend, but no type section was designated by Furnival.

The upper boundary of the Frenchman is the lowest commerical and mappable coal seam, which is placed in the Ravenscrag Formation. In the Eastend area this lignite seam

is called the No. 1 or Ferris seam. It is several feet thick and has a break of carbonaceous shale of about 6 inches separating a lower thicker bed from an upper. In Montana, where the Hell Creek is the correlative of the Frenchman and the Fort Union of the Ravenscrag, the lowest mappable lignite is also taken as the base of the Fort Union (Colton, 1955).

The lower boundary of the Frenchman is marked by a surface of erosion as McLearn had already recognized (*in* Fraser *et al.*, 1935, p. 37). Where the pre-Frenchman erosion is nil or very slight, the Frenchman Formation rests on the Battle. In other places the Frenchman rests on one of the three members of the Whitemud, the Eastend, or the Bearpaw. The greatest amount of erosion took place in the northern and eastern parts of the area on the flanks of the Cypress Hills, where as much as 200 feet of older formations was removed before deposition of the Frenchman. Less material was removed from the center of the hills in the Ravenscrag-Eastend area. Here the Frenchman commonly rests on Battle or on some stratum of the Whitemud Formation and has some local channelling down to the Eastend sands. To the east the Frenchman lies on progressively lower strata down to some part of the Bearpaw shale at the eastern boundary of the map area. On the northern slope of the Cypress Hills also, the Frenchman overlies the Bearpaw.

Distribution and Thickness

Plate 1 shows the distribution of the Frenchman Formation in the eastern Cypress Hills. On older geological maps covering this general region (Fraser *et al.*, 1935, Regina sheet) the Frenchman was included in the Ravenscrag. In general the Frenchman on Plate 1 coincides with the outer, lower border of the Ravenscrag as mapped on the Regina sheet. Important differences are noticeable in the northeastern sector of Plate 1 where large areas previously mapped as Eastend are now shown as Frenchman. These are all outside the area examined by McLearn, as he did not map north of T. 8 (Fraser *et al.*, 1935, p. 2). Key outcrops for this reinterpretation occur along the eastern bank of Swift Current Creek in NW ¼ sec. 13, T. 9, R. 20, W3., where the writer found typical Frenchman fossils in the sand unit (Table 3). These banks are mapped as Eastend on the Regina sheet (Fraser *et al.*, 1935, map) and no Ravenscrag (which would include the Frenchman) is shown east of the creek. From these key outcrops the Frenchman sand can be followed north where other fossil localities verify in places the lithologic correlations. Plate 1 does not show the distribution of the Eastend, but it can be seen from the abbreviated sections that no Eastend beds are present north of T. 8, where the Frenchman everywhere lies directly on Bearpaw.

The thickness of the Frenchman varies with the depth of erosion that took place before the formation was deposited. It ranges from a minimum of 28 feet to more than 222 feet where the Frenchman rests on some horizon in the Bearpaw. The Frenchman increases in thickness from the town of Eastend to the north and to the east.

Lithology

Description of Two Frenchman Lithostratigraphic Units

Two lithostratigraphic units make up the Frenchman Formation. One is here referred to as the sand lithosome, the other as the clay lithosome. These units were previously referred to as zones (McLearn, *in* Fraser *et al.*, 1935, p. 41), phases (Russell, 1948, p. 32), and facies (Kupsch, 1956, p. 20). The increased knowledge of the stratigraphical relationships between the two lithological units of the Frenchman and the recent redefinition and clarification of many lithostratigraphical terms seem to justify the adoption of the term lithosome. Wheeler and Mallory (1956, p. 2718–2719) define a lithosome as a lithostratigraphic body which is mutually intertongued with one or more bodies of differing lithic constitution. They point out that several writers previously referred to such vertico-laterally delineated units as "facies," but that the restricted definition of lithofacies as laterally segregated, statistical, lithic variants of a stratigraphic interval necessitates the introduction of the term lithosome.

The sand unit of the Frenchman Formation consists of medium- to fine-grained "salt-and-pepper" sand or loosely cemented sandstone, which locally may be well cemented into large

TABLE 2.—LITHOLOGICAL CHARACTERISTICS OF EASTEND AND FRENCHMAN SANDS

Sediments / Lithological characteristics	Sand from	
	Sand lithosome of Frenchman Formation	Eastend Formation
Color	Somber	Pale
	Light olive gray to dark greenish gray	Grayish yellow to light dusky yellow
	In places yellow to brown	Some orange siltstone
Grain size	Medium- to fine-grained	Very fine-grained
	Low silt and clay content	High silt and clay content
Sorting	Poorly sorted	Well sorted
Presence of chert	Common	Present
	Visible	Microscopic
	"Salt and pepper" sand	
Presence of carbonates	Unindurated sand generally noncalcareous	Commonly calcareous
	Indurated masses calcareous	
Presence of mica	Mica absent or subordinate	Commonly micaceous
Presence of carbonaceous material	Fragments common	Comminuted plant material scattered through sediment
	Some thin lignite seams	
	Large pieces of lignitic fossil wood in conglomerates	
	Carbonaceous shales in places	
Heavy minerals	Unstable heavy minerals predominate	Stable heavy minerals predominate
	Green hornblende prominent (40%)	Green hornblende present (3%)
Stratification	Very thick-bedded	Thin-bedded
Cross-stratification	Common	Present in places
	Medium to large scale	Small scale
Induration	Indurated parts common	Few indurated strata
	Large blocks of indurated calcareous sand ("loglike concretions") commonly lying on outcrop	
Splitting property	Indurated parts massive	Indurated strata flaggy
Fossils	Fairly common	Rare
	Nonmarine; dinosaur bone fragments, etc.	Marine (in eastern Cypress Hills); pelecypods, etc.
Associated sediments	Some clay lenses in places	Clay lenses and beds common
	Locally clay pellets	In places interbedding of sands and clays
Lower contact of Eastend and Frenchman formations	Conglomerates of clay pebbles with few quartzite pebbles common at or near lower contact of Frenchman where erosional disconformity developed	No conglomerates observed in Eastend
		Contact between Bearpaw and Eastend gradational

hard sandstone masses. The sand is light-olive gray to dark-greenish gray, in places yellowish orange to brown. Medium-scale cross-stratification is a common feature. Large indurated masses, well cemented by calcium carbonate, are rounded and elongated parallel to the bedding. They are commonly referred to as "loglike concretions." In places lenses or thin beds of gray silt and clay occur in the coarser sandstone. Carbonaceous material, such as small macerated plant fragments, is common in the sand. Locally thin beds of carbonaceous shale, blocks of lignitic fossil wood, and thin stringers of lignite are found. The sediment can best be described as a subgraywacke of the "salt-and-pepper" variety.

In some outcrops it is not easy to distinguish the sand unit of the Frenchman from the sands of the underlying Eastend Formation. It is important to discriminate between them where

the Frenchman lies on the Eastend or on the Bearpaw. Lithological characteristics helpful in distinguishing Frenchman sand from Eastend sand are listed in Table 2. Singly, any of the criteria in Table 2 may not be diagnostic, but in combination they are sign posts for the presence of either formation in areas where the stratigraphic position of the questionable sand is obscure because of the erosional removal of the characteristic Whitemud beds separating Eastend from Frenchman. McLearn (*in* Fraser *et al.*, 1935, p. 25) was aware of this problem when he wrote: "Owing to difficulty of identification of the Eastend in areas where the Whitemud is absent a sand at the base of the Ravenscrag (= Frenchman) may be mistaken for Eastend and vice versa. This may be true even of the most painstaking work if the specific criteria are obscure." Therefore an extended list of lithologic characteristics (Table 2) may be helpful. Some of these characteristics, such as grain size, lignitic material, and nature of contact were used by McLearn; others were discovered by petrological investigations of samples conducted by Fraser; others represent field and laboratory observations by the writer. Although the Frenchman fossils do not necessarily require a time interval younger than Eastend, they are helpful in distinguishing Frenchman sands from Eastend sands: fossils are scarce in the Eastend, and all those in that formation, within the map area, are marine (Russell, 1951, p. 26–27), whereas a continental origin is indicated for the Frenchman by both animal and plant fossils.

The clay unit of the Frenchman, different from the sand unit in its gross aspects, is composed of grayish, grayish-green, and bluish to purplish bentonitic clays. Its physiographic expression is in sharp contrast to that of the cliff-forming, or grass-covered, sand unit. The clay unit is characteristically exposed in rounded badland outcrops with a cracked surface and a sparse or absent vegetation cover. Some sand lenses are in places incorporated within the clay unit. They are very fine-grained and would not be easily distinguishable from the Eastend sand except that they are thin and are interbedded with clay strata characteristic of the clay unit of the Frenchman Formation.

Relationship of the Two Lithostratigraphic Units

The relationship between the sand and the clay units of the Frenchman has been variously interpreted by previous workers. McLearn (*in* Fraser *et al.*, 1935, p. 41) considered the sand as the basal unit of the Lower Ravenscrag (= Frenchman) and indicative of some erosion preceding the deposition of the clay unit. He believed the sands to be absent where the entire Whitemud (including Battle) is present, which indicates little or no erosion before Frenchman time. In addition to this basal sand he mentioned the presence of massive sands, in the middle and near the top of the Frenchman near Climax road bridge, with clays between the three sand units.

Furnival (1946, p. 95) regarded the sands as the typical Frenchman and described the clay unit as intercalated with the thick sandstone lenses and commonly occurring at the top of the formation. Furnival followed McLearn in regarding the sand as the basal unit of the Frenchman Formation. He drew the base of the Frenchman at the base of the lowest sand and referred any clays under this sand and overlying the Whitemud to the Battle Formation.

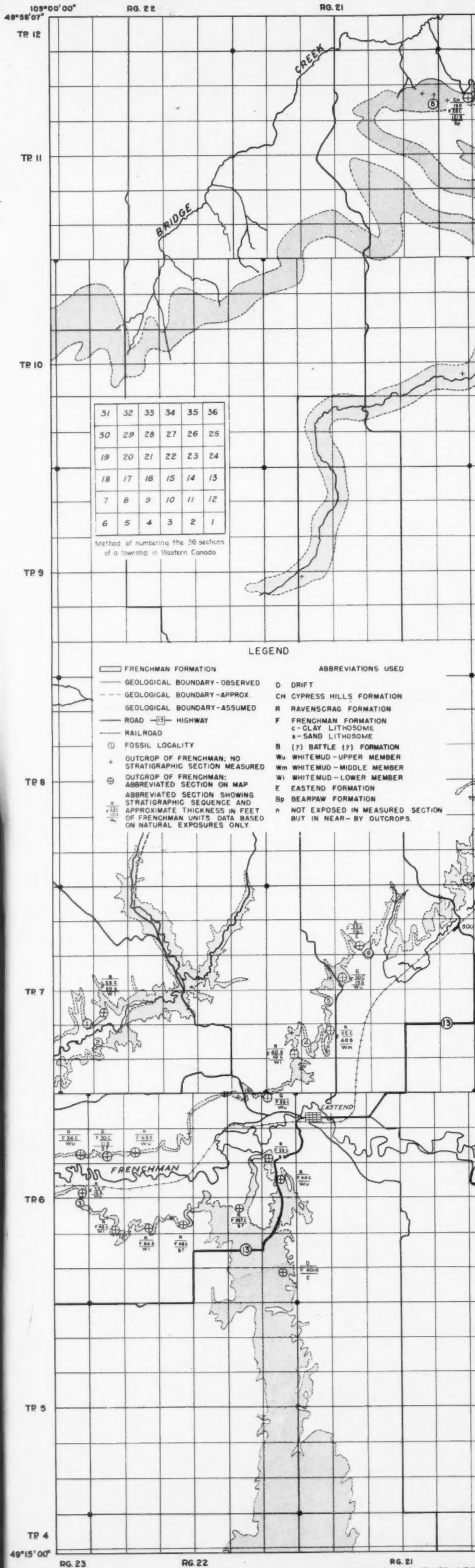
Russell (1951, p. 34) indicated that the clay unit represents the typical Frenchman and that the sand occurs at the top of the formation. He regarded the sand as local channel filling at the beginning of Ravenscrag deposition. He pointed out that equivalence of the sand and clay units was not proved and that a local unconformity might be present between the Frenchman and the Ravenscrag. Russell suggested removal of the sand unit from the Frenchman and regarded it as the lowest unit of the Ravenscrag.

In an attempt to solve the relationships between the two lithostratigraphic units of the Frenchman the writer analyzed the lithological characteristics of 49 sections in the eastern Cypress Hills. Plate 1 shows: (1) a complete Frenchman section may be composed wholly of the sand unit, wholly of the clay unit, or of both; (2) locally either the clay or the sand is the lowest unit, whether much erosion preceded Frenchman deposition or not; (3) either clay or sand may be the upper unit; (4) the two units may alternate vertically; (5) the sand unit is not invariably present at the base of the French-

TABLE 3.—FOSSILS FROM FRENCHMAN FORMATION, EASTERN CYPRESS HILLS

Locality			Lithosome in which fossils found		Fossils
Number on Plate 1	Name	Description (Ranges west of third m.)	Sand	Clay	
1	Topham's ranch	NE Sec. 7, T.7, R.22	×		Crest, horn, and vertebrae of *Triceratops;* collected by H. Jones
					Two vertebral centra of *Champsosaurus*
					Crocodilian scute
					Fragments of *Aspideretes* shells
2	Little Frenchman Creek	SE Sec. 7, T.7, R.22	×		*Carpolithus (Cycadinocarpus?) ceratops;* determined by C. A. Arnold.
3	Knollys	SE Sec. 13, T.6, R.23		×?	Skull of *Triceratops;* collected by C. M. Sternberg
4	Eastend	SW Sec. 8, T.7, R.21	×		Crocodilian tooth
					Costal plate of *Aspideretes* sp.
					Amphibian mandible
					Turtle vertebrae
5	Chimney Coulee	NE Sec. 17, T.7, R.21	×?		Supraorbital horn core of *Triceratops;* collected by H. Jones
6	Anxiety Butte	SE Sec. 28, T.7, R.21	×		Bone fragments, possibly dinosaurian; determined by L. S. Russell
					Shell fragments of hard- and soft-shelled turtles; determined by L. S. Russell
					Fish vertebrae; determined by L. S. Russell
					Scales of *Lepidosteus;* determined by L. S. Russell
					Mandibular tooth of a large ceratopsian
7	Swift Current Creek	NW Sec. 13, T.9, R.20	×		Fragments of *Basilemys* and *Aspideretes* shells
					Portions of an anterior caudal vertebra of a large carnivorous dinosaur, about the size of *Tyrannosaurus rex*
8	Lookout Hill	SE Sec. 25, T.11, R.21		×	Claw from fore foot of a carnivorous dinosaur
					Fragments of *Aspideretes* shell
9	Northwest of Busse's ranch	NE Sec. 17, T.5, R.19		×	Crocodilian scute
					Fragments of turtle (*?Aspideretes*) shell
					Dorsal vertebral centrum of *Champsosaurus*
10	Northeast of Busse's ranch	NE Sec. 16, T.5, R.19		×	Skull of *Triceratops;* collected by H. Jones
					Fragment of carnivorous dinosaur bone
					Dermal scute or cranial fragment of an armored dinosaur
					Fragment of right-angular bone (mandible) of a small crocodilian
					Fragmentary crocodilian scutes and cranial elements
					Crocodilian tooth
					Large ceratopsian tooth
					Fragment of a large tooth, possibly crocodilian
					Small carnivorous dinosaur tooth of a form recognized in Alberta, but unnamed

If not indicated otherwise, all fossils collected by W. O. Kupsch; determinations by W. Langston in collaboration with L. S. Russell and C. M. Sternberg



DISTRIBUTION OF FRENCHMAN FORMATION SASKATCHEWAN.

RG. 20 RG. 19 106°00'00" 49°56'07"

TP. 12

TP. 11

TP. 10

TP. 9

TP. 8

TP. 7

TP. 6

TP. 5

TP. 4

49°15'00"

RG. 20 RG. 19 106°00'00"

MERIDIAN

ON IN THE EASTERN CYPRESS HILLS,
N, CANADA

man Formation where pre-Frenchman erosion has been considerable; (6) clay beds of the Frenchman are locally in direct contact with Bearpaw strata; (7) there is no apparent relationship between the development of either lithostratigraphic unit and the total thickness of the Frenchman in an outcrop.

Because of these relationships the writer proposes that the sand and clay units are lithosomes of equal importance in the Frenchman Formation. These mutually intertongued bodies are not to be regarded as members of the Frenchman because they do not produce any definite stratigraphic sequence over more than a restricted and therefore unmappable area.

Where the disconformity at the base of the Frenchman is well defined, as in many outcrops along the eastern and northern rim of the Cypress Hills where the Frenchman lies on strata that are low in the stratigraphic section, it is in several places characterized by a conglomeratic stratum. The conglomerate consists predominantly of rusty siltstone pebbles, with some fossil wood, dinosaur bones, and other fragmentary fossils in places. A few quartzite pebbles are commonly present; they closely resemble the quartzite pebbles of the later Cypress Hills Formation. This basal conglomerate is apparently of wide occurrence, although it is discontinuous. It has been reported from the Wolf Point area, Montana (Colton, 1955).

No indications of a disconformity at the top of the Frenchman exist anywhere in the eastern Cypress Hills area. The lithological change from the Frenchman sediments to the bottom coal bed of the Ravenscrag is abrupt but conformable. A thin clay bed underlies the coal, even where the sand lithosome is the highest unit in the Frenchman.

Paleontology

The Frenchman beds in the eastern Cypress Hills have yielded many dinosaurian bones in addition to remains of turtles, fishes, crocodiles, and some fossil fruits (Table 3). An upper Cretaceous age is indicated by the fossils, especially by the dinosaurs.

An attempt was made to determine the lithostratigraphical unit of the Frenchman from which the fossils were collected; location and lithology are given in the literature (Fraser *et al.*, 1935, p. 42) for some fossils collected by Sternberg and Jones in the eastern Cypress Hills area. Additional information was kindly given by Mr. H. Jones of Eastend. For all fossils collected by the writer the lithology of the Frenchman in which they were found was noted. It is evident from Table 3 that similar faunules are present in both the clay and sand units. Previously Russell (1951, p. 33) had stated that no fossil bones, diagnostic or otherwise, have been found in the sand unit and that correlation based upon fossils should therefore be applied only to the clay unit. The finding of similar fossils in the clay and sand units supports the inference that the two units are equivalent mutually intertongued bodies.

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Manuscript Received by the Secretary of the Society, April 13, 1956

Published by Permission of the Director, Saskatchewan Geological Survey

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 68, PP. 421–444, 17 FIGS., 2 PLS. APRIL 1957

ORIGIN OF MARINE-TERRACE DEPOSITS IN THE SANTA CRUZ AREA, CALIFORNIA

BY WILLIAM C. BRADLEY

ABSTRACT

The sea floor in the Santa Cruz area, California, is a modern wave-cut platform carved with 2–4 feet of relief into the Miocene Monterey Formation. The platform slopes seaward at an angle of less than 1° and is slightly concave upward within half a mile of the sea cliff. It is either bare or mantled by less than 4 feet of sand whose grain size tends to be less than that of modern beach sand and appears to decrease seaward.

The lowest of five prominent marine terraces in the same area, here called the 100-foot terrace, contains an old wave-cut platform which is similar to the modern platform and probably was developed in the same manner. Overlying this old platform are marine deposits which contrast with the modern platform deposits, but which resemble beach deposits in that their grain size is similar, their thickness is 5–15 feet, and they contain laminated structures that include black sands. Rock-boring mollusk holes in the old platform, now buried beneath the terrace deposits, suggest that these deposits were laid down after the platform was cut. Pyroxene, sphene, and shells have been leached from the terrace deposits to a progressively greater extent toward the old sea cliff; this leaching suggests that the deposits are oldest near the cliff and become progressively younger toward their distal portions. The marine-terrace deposits are interpreted as beach deposits which were prograded seaward across the platform during a period of slow emergence.

The marine-terrace deposits consist of two types of sand: a coarse mica-poor variety similar to beach sand, and a fine mica-rich variety similar to some neritic sediments in Monterey Bay. The coarse sand is interpreted as normal beach sand, derived from streams and redistributed along the prograding beach by longshore drift. The fine sand is interpreted as former neritic material which was reworked during emergence into the prograding beach where it was interlaminated and mixed with the other sand.

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Introduction

Marine terraces along the west coast of North America characteristically consist of a cut bedrock platform mantled by deposits that are marine or nonmarine, or both, in origin. Genesis of the terrace deposits, particularly the marine, has long been a problem. In an attempt to achieve a better understanding of the origin of such deposits, an investigation was made of the well-preserved marine terraces in the Santa Crus area, California (Fig. 1). The Santa Cruz area, arbitrarily restricted to the coastal area between Santa Cruz and Waddell Creek, was chosen for study because available submarine data on bottom sediments (Galliher, 1932; 1935) and on a modern bedrock platform provided a standard to which analogous features in the marine terraces could be compared. This paper, in presenting some results of the investigation, centers around the sea floor and lowest prominent terrace. Higher terraces are considered briefly, but are so similar to the lowest one that a full discussion of the entire sequence would not add significantly to the paper.

Acknowledgments

The marine-terrace study was supervised by A. D. Howard, Stanford University, and the writer is pleased to acknowledge his aid and interest throughout and his helpful criticism of this paper. Special thanks for numerous fruitful discussions are also due G. A. Thompson, Stanford University, and Sheldon Judson, Princeton University. R. E. Garrison and T. C. Silver assisted with surveying, and R. A. Sands, D. E. Feltham, R. L. Christiansen, and D. E. Lee helped collect bottom and beach samples. A. D. Russell, Santa Cruz city engineer, and Danta Falcioni, engineer for the Santa Cruz Portland Cement Company at Davenport, made available some of the submarine data used in this paper. Grants from the Shell Fund for Fundamental Research at Stanford University supported the study.

Geologic Setting

Marine terraces in the Santa Cruz area occupy the lower seaward slopes of Ben Lomond Mountain, a northwestward-trending ridge separated from the rest of the Santa Cruz Mountains by the San Lorenzo River lowland. Ben Lomond is asymmetric in cross profile, steep toward the San Lorenzo River on the northeast and gentle toward the Pacific Ocean on the south and southwest. This form suggested to Branner *et al.* (1909, p. 10) and Willis (1925, p. 647) that the mountain is a block which has been faulted on the northeast and tilted toward the southwest. Rode (1930, p.

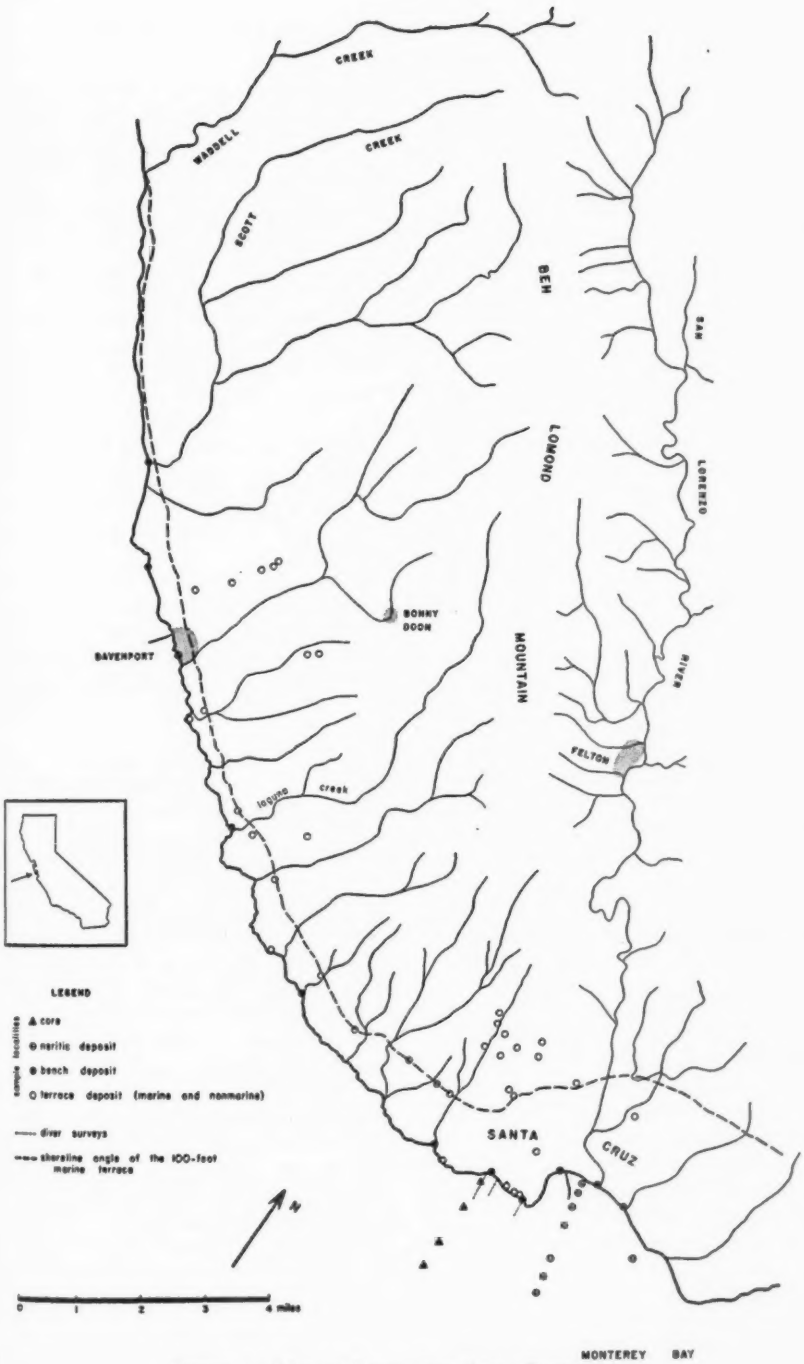

FIGURE 1.—MAP OF SANTA CRUZ AREA, CALIFORNIA

21–28), however, concluded that faulting has been of minor importance, and that the mountain is essentially an asymmetric elongate dome.

The core of Ben Lomond Mountain, consisting of crystalline schists and marbles intruded by quartz diorite (Fitch, 1931), is flanked by generally outward-dipping Miocene sedimentary formations: Santa Margarita, Monterey, and Vaqueros (Branner *et al.*, 1909; Rode, 1930; Fitch, 1931; Page and Holmes, 1945). Inasmuch as the marine terraces are developed principally in the Monterey Formation, only this unit will be described here briefly; Bramlette's (1946) full discussion of the Monterey is applicable to the rocks of the Santa Cruz area. Locally, as along part of West Cliff Drive in Santa Cruz, the formation is a mudstone, but the most common rock types are jointed, thin-bedded porcelanites, porcelanous shales, and shales which are dark gray or brown if fresh and white if weathered. On the southwest side of Ben Lomond Mountain the Monterey dips generally seaward at angles that, except for local variations, range from 3° to 15°.

Marine terraces give the seaward side of Ben Lomond the appearance of a crude gigantic staircase. Their exact number and elevations have not been satisfactorily determined, although it is generally agreed that there are five prominent, and several indistinct, terraces between sea level and 850 feet. The terraces have been described by Lawson (1893), Wilson (1907), Branner *et al.* (1909), Rode (1930), Page and Holmes (1945), and Alexander (1953).

Nomenclature

Figure 2 is a diagrammatic cross section of a wave-cut platform and a marine terrace. The bedrock part of the sea floor, with or without associated deposits, is the wave-cut platform, a term synonymous with "wave-cut bench" and "wave-cut terrace", and the junction between platform and sea cliff is the shoreline angle (Davis, 1933, p. 1051), a term synonymous with "back edge" of Rode (1930, p. 29) and Alexander (1953, p. 11). The precise relationship between the shoreline angle and sea level is controversial, but for the present discussion it is assumed that the shoreline angle marks the approximate level of high tide Kuenen, 1950, p. 538). Beach deposits commonly bury the shoreline angle. Using Thompson's (1937, p. 725–726) terminology, the beach deposits occurring between the low-tide limit and the upper limit of permanent saturation are the lower foreshore deposits, those occurring between the upper limit of permanent saturation and the crest of the beach are the upper foreshore deposits, and those occurring behind the crest are the back-shore deposits.

In a zone near shore, sea water moves in a current pattern which Shepard and Inman (1950, p. 200–201; 1951, p. 52) have called the "nearshore system" of currents, and this zone is here called the nearshore zone. The seaward margin of the nearshore zone is the line where shoaling water begins to appreciably affect wave motion; the inner portion of the zone, lying between breakers and beach, is the surf zone.

When a wave-cut platform is raised above sea level and cliffed by wave action it becomes a marine terrace. The old platform and shore-line angle are commonly mantled by deposits which are either marine or nonmarine, or both, in origin.

Methods of Study

Submarine Data

Data on the modern wave-cut platform comes from cores, from bottom surveys made by divers (Hyde and Howe, 1925–1926), and from the construction of two piers, one at Santa Cruz (A. D. Russell, personal communication) the other at Davenport (Helen, 1935; Santa Cruz Portland Cement Company, Davenport). The locations of cores, bottom surveys, and piers are shown in Figure 1.

Data on bottom sediments came from Galliher (1932; 1935) and from 10 samples collected at depths of 100 feet or less. One of the 10 samples was collected by Galliher (Galliher, 1935, sample 115); two others were cores; the rest were collected by skin divers. The location of these samples is shown in Figure 1. These and all other sedimentary samples were examined in the laboratory.

Beach Data

Twenty-three beach samples, some from single laminae, some channel samples, were

collected between Santa Cruz and Scott Creek (Fig. 1); lower foreshore, upper foreshore, and back-shore deposits from both exposed and protected beaches are represented. Although sampling did not follow a definite grid pattern, the writer believes that for the purposes of the present study these samples are representative of beaches in the Santa Cruz area.

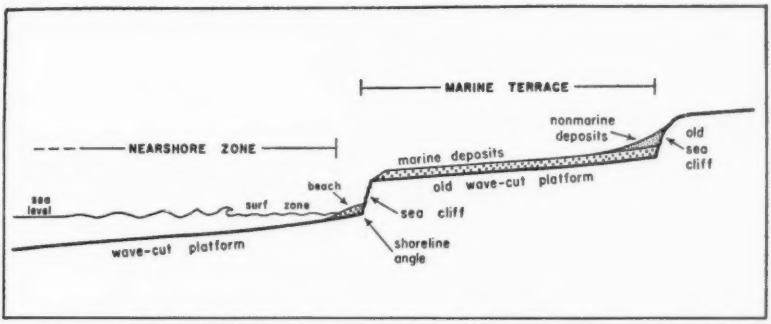

FIGURE 2.—DIAGRAMMATIC CROSS SECTION SHOWING TERMINOLOGY USED IN THIS PAPER

Terrace Data

Former wave-cut platforms in the marine terraces were observed and surveyed with an engineer's level and rod. Because the various platforms are similar, only that belonging to the lowest prominent terrace is described here.

Forty-two marine and 16 nonmarine samples from the deposits of the lowest terrace, and 32 from higher terraces, were collected at the modern sea cliff and in road and quarry cuts between Santa Cruz and Davenport (Fig. 1). Because the lowest terrace narrows to the northwest from Santa Cruz (Fig. 1), road cuts and the present sea cliff expose a fairly continuous section of the terrace deposits extending seaward 1½ miles from the former sea cliff. These deposits are so weathered that they lie within the A, B, and C horizons of the soil profile, and in some places within the A and B horizons. At one locality along West Cliff Drive in Santa Cruz the deposits were sampled at approximate 1-foot intervals from top to bottom. Aside from the leaching of unstable minerals within the A and B horizons, no systematic variations in the mineralogy were detected within the vertical range of the 12-foot deposits. Samples from this terrace, all collected from the C horizon, may therefore reasonably be compared with one another.

Laboratory Procedures

Sedimentary samples were sieved and the results represented on histograms and cumulative curves, from which median diameters and sorting indices were calculated. The content of mica, shells, and fragments of locally derived rocks were determined with a binocular microscope. Mica, if present, is concentrated in the coarser fractions of any sand (Fig. 3), as has been reported elsewhere by Emery *et al.* (1952, p. 542), Hutton (1952, p. 27), and Inman (1953, p. 37). Heavy minerals in the samples tend to be concentrated in the 0.147- to 0.074-mm sand fraction (Fig. 4), and these were separated from

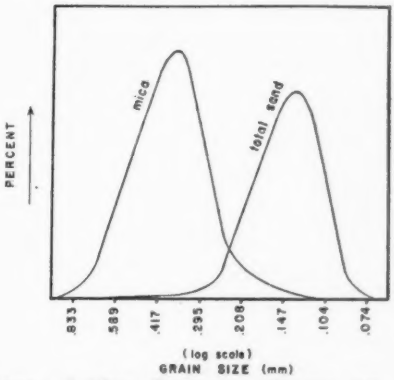

FIGURE 3.—SIZE FREQUENCY CURVES FOR A MARINE-TERRACE SAND AND ITS CONTAINED MICA

Mica is concentrated in the coarser sand fractions.

the light minerals in Bromoform. The 0.147- to 0.074-mm fraction was caught on two screens, no. 150 and no. 200 (Tyler numbers). Of these two, the fraction caught on the no. 150 screen (0.147-0.104 mm) proved to be most useful, and

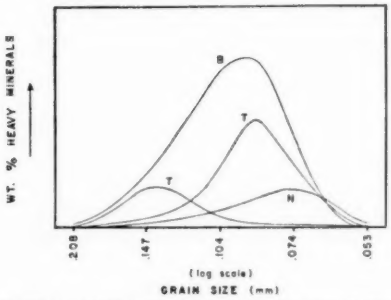

FIGURE 4.—SIZE FREQUENCY CURVES FOR HEAVY MINERALS OF FOUR SAMPLES THAT ILLUSTRATE USUAL RANGE IN THE CURVES

B is beach sand, *N* is neritic sand, *T* is marine terrace sand. Heavy minerals tend to be concentrated between 0.147 and 0.074 millimeters.

consequently only the heavies from this fraction were studied in detail. Therefore, the mineral percentages discussed in this paper are not indicative of absolute percentages in the samples but are used solely to indicate comparative variations between different sands.

Heavy minerals were mounted on glass slides and percentages determined by counting between 250 and 350 grains per slide with a mechanical stage. Pyroxene and sphene, relatively abundant within the 0.147- to 0.104-mm size range, were especially useful in comparing different sands. Shallow neritic, beach, and terrace deposits contain a heavy mineral suite similar to that of beaches along this part of the california coast reported in detail by Hutton (1952). The important and accessory minerals found in neritic, beach, and terrace deposits in the Santa Cruz area are:

Important minerals	*Accessory minerals*
Magnetite	Tremolite-actinolite
Ilmenite	Titanaugite
Hornblende	Epidote
Augite	Sphene
Hypersthene	Zircon

Oxyhornblende, glaucophane, garnet, biotite, tourmaline, apatite, zoisite, and others commonly occur in trace amounts.

MODERN WAVE-CUT PLATFORM

Platform

The sea floor in the Santa Cruz area is a modern wave-cut platform carved into the Monterey Formation. Figure 5 shows the profile

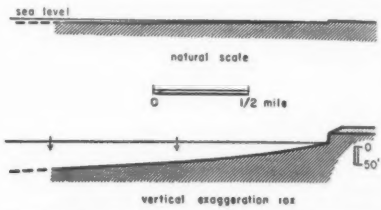

FIGURE 5.—PROFILE OF MODERN WAVE-CUT PLATFORM AT SANTA CRUZ

Beach is not shown. Arrows indicate where cores were collected.

of the platform at Santa Cruz; the known profile at Davenport is similar and is not reproduced here. The inner half mile of the platform, examined by divers (Hyde and Howe, 1925–1926, p. 259–260), is continuous; the rest of the profile, to a distance of about 1½ miles from the sea cliff, is drawn on the basis of two bedrock cores (Fig. 5). Thus the modern platform at Santa Cruz is at least 1½ miles wide and extends to a depth of at least 75 feet. How much farther seaward it continues is not known; a core taken from a depth of 102 feet consisted of 14 inches of sand, but it did not penetrate deeply enough to show whether the sand was a thin sheet overlying the bedrock platform.

The platform slopes seaward at a low angle and is slightly concave upward; the concavity is most pronounced within half a mile of the sea cliff. The gradient of the platform, about 1° near the shoreline angle, decreases seaward to about 0.5° where it becomes nearly uniform. In this area the Monterey Formation dips seaward at about 6° and its truncation by the platform proves that the latter is an erosional feature.

In drawing the profile in Figure 5 all irregularities of less than 5 feet were evened out, and the illusion that the platform is perfectly smooth is created. Actually, the platform at Santa Cruz has a natural relief which is as great as 7 feet (Hyde and Howe, 1925–1926, p. 259–260), but which is commonly between 2 and 4 feet. At Davenport the platform relief is generally 2–4 feet but locally is as much as 10

feet (profile, Santa Cruz Portland Cement Company, Davenport). The local relief at Santa Cruz is related in part to boulders, which are probably calcareous concretions etched out of the Monterey Formation, and in part to bedrock ridges, which are probably low cuestas.

the construction of piers at Santa Cruz (A. D. Russell, personal communication) and at Davenport (profile, Santa Cruz Portland Cement Company, Davenport; Helen, 1935), and from diver surveys (Hyde and Howe, 1925-1926); data beyond half a mile from shore come

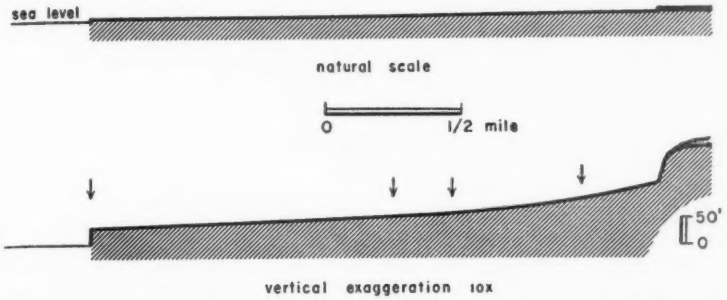

FIGURE 6.—PROFILE OF OLD WAVE-CUT PLATFORM OF 100-FOOT MARINE TERRACE AT SANTA CRUZ
Terrace deposits are not shown. Arrows indicate individually surveyed points.

The profile has been drawn to represent a single continuous platform, even though only the inner half mile is known to be continuous (at both Santa Cruz and Davenport); the rest of the profile is based on two cores (Fig. 5). It is possible, therefore, that more than one platform is present, separated by one or more low cliffs. Fairbridge (1952, p. 348) reported that submerged marine terraces are common features on continental shelves; Stearns (1945) and Emery *et al.* (1954) reported submerged terraces bordering some of the Pacific islands; Newell and Imbrie (1955, p. 7) recorded similar features on the Bahama Banks; Upson (1949, p. 107–108), Dietz and Menard (1951, p. 2013–2014), Holzman (1952, p. 100), and Emery (1954, p. 107) reported terraces on the southern California shelf. Similar features could exist in the Santa Cruz area but may not have been detected because of the wide spacing of cores, the lack of sensitive soundings, and the natural relief of the platform. On the basis of the data available at present, however, the modern platform is believed to have a profile as shown in Figure 5.

Platform Deposits

Thickness.—Available data indicate that the modern platform is either bare or mantled by less than 4 feet of sand. Data on the platform deposits within half a mile of shore come from two cores consisting of bedrock with no sand veneer.

The slight thickness of the platform deposits does not appear to be a local or a seasonal effect. Construction of a sewer outfall line during the summer of 1954 revealed that the bottom was in essentially the same condition as it was when examined by divers in the winter of 1925–26 (A. D. Russell, personal communication). In addition, the city of Santa Cruz keeps a continuous depth record at its municipal pier, and Russell reports that beyond the submarine extension of the beach the thickness of sand fluctuates generally less than a foot throughout the year.

Grain size.—Galliher (1932; 1935) found that Monterey Bay sediments become progressively finer-grained seaward, but that the grain-size zones only roughly parallel the isopath lines. Only the inner portion of the sediments studied by Galliher relate to the modern platform, and in this narrow zone there is at best only a rough relationship between grain size and depth. The grain size of neritic samples and the range in grain size of beach samples are shown in Figure 7. The dashed line indicates that except for two samples there is a slight but general seaward decrease in grain size, although because of lack of sufficient samples this decrease must remain only a suggestion. The solid line indicates that neritic sediments tend to be finer-grained than beach sediments, and this is supported by

Galliher's observations. The coarse neritic sample from 30 feet depth was collected by Galliher, and the reason for its unusual size is unknown. The coarse sample from 50 feet was collected at the top of a sand ridge that rose 1 or 2 feet above the adjacent sea floor. Shepard and Wrath (1937, p. 43), Clements and Dana (1944, p. 354), Emery (1952, p. 1106), and Wimberley (1955, p. 33) interpreted coarse sediments on the continental shelf in southern California as relict beach or shallow-water deposits laid down when sea level was lower than it is today. Possibly the low sand ridge at Santa Cruz is a submerged beach ridge.

FIGURE 7.—RELATIONSHIP BETWEEN DEPTH AND GRAIN SIZE FOR NERITIC SAMPLES FROM SANTA CRUZ AREA

Coarsest and finest beach samples are included for comparison. Solid line is at 0.200 millimeters. Dashed line suggests a rough seaward decrease in grain size.

Mapping in the last 25 years has shown that sediments on the continental shelf are highly irregular in grain size. As Emery (1952, p. 1107) pointed out, however, few attempts have been made to distinguish between the different types of sediments. Emery (1952) distinguished five types of sediments on the southern California shelf: authigenic, organic, residual, relict, and detrital; only detrital sediments are being carried out from shore today. A map of these deposits yields a confused pattern of grain sizes, but when other varieties are eliminated the detrital class does show a general seaward decrease in grain size (Emery, 1952, p. 1107; 1954, p. 109). Possibly a similar mixture of different sediments exists in the Santa Cruz area, and is at least partly responsible for the poor correlation between depth and grain size shown in Figure 7.

Significance of Platform Concavity

The slope of the modern platform seaward suggests that it is related to the transportation of sediment, and the occurrence within shallow neritic sands of subaerially-etched pyroxene grains (Fig. 10) indicates that there is at least a seaward component of sediment movement on the sea floor. The modern platform is concave within half a mile of the shore, where it coincides with the nearshore zone.[1] Johnson (1919, p. 211) and Kuenen (1950, p. 303) correlated the seaward decrease in platform gradient with a seaward decrease in the size and amount of load moving across the platform, the belief being that a diminishing load can be moved on a diminishing gradient. They felt that the reduction in load was due to abrasion of particles and removal of the fine attrition products in suspension.

What influence grain size of load has on the platform gradient is in doubt because there is only a suggestion that it decreases seaward (Fig. 7). On the other hand, quantity of load in the nearshore zone does decrease seaward from the shore. This is partly the result of attrition, and partly the result of natural sorting processes in the nearshore zone that selectively move fine material seaward and leave coarser material to drift along within the zone. The sorting comes about because the short sharp onshore movement of water can transport coarser particles, whereas the slower offshore movement can transport only finer particles, particularly those in suspension (Shepard and LaFond, 1940, p. 284; Grant, 1943, p. 118–120).[2]

[1] Shepard and Inman (1950, p. 202) state that at times the near-shore zone may extend out to distances of several thousand feet from shore.

[2] The short sharp onshore movement is a result of the greater onshore orbital velocity of water particles; the slower offshore movement is the result of gravity and the slower offshore orbital velocity.

Eaton (1951, p. 143) refers to a certain grain size that is in equilibrium with prevailing conditions and stays at a constant depth of water, at the same time that coarser particles move onshore and finer ones offshore. Rip currents assist in sorting by transporting fine material seaward (Shepard and LaFond, 1940, p. 284).

If the slope of the modern platform is related to sediment movement, it seems reasonable to correlate the platform concavity with a seaward decrease in amount and perhaps size of load within the nearshore zone where abrasion and sorting are most effective.

Platform as a Profile of Equilibrium

According to the concept of the submarine profile of equilibrium, as conceived by Gilbert (1885, p. 83; 1890, p. 33–34), Fenneman (1902, p. 22–32), Davis (1909, p. 701–703), Johnson (1919, p. 211–224), Barrell (1920, p. 352–358), Twenhofel (1939, p. 213–215), and Kuenen (1950, p. 302–306), the condition of equilibrium, an ideal which the natural system constantly strives to attain, exists where the ocean bottom descends with such declivity that the agents of transportation can just disperse whatever load is present within the system. The platform gradient is thus the flattest one that allows transportation. The direction of load movement was considered to be a resultant of two components: movement seaward with undertow and movement parallel to the shore by longshore drift.

It is now known that the nearshore movement of debris is more complicated than any simple resultant of longshore and offshore motions. Shepard and Inman (1950, p. 200–202; 1951, p. 50–52) and Inman (1954, p. 31) found that water within the nearshore zone circulates in a complex cell-like pattern of rotational movements around vertical axes. Briefly, the components of this circulation pattern are: (1) the onshore movement of water as mass transport outside and surf itself inside the surf zone[3]; (2) the seaward return of water as undertow and rip currents; (3) longshore currents, partly the longshore component of mass transport and undertow, and partly the result of oblique wave approach within the surf zone and the expanding heads of rip currents outside the surf zone. Shepard and Inman (1951, p. 50) believe that this circulation pattern is common along coasts; rip currents observed by this writer suggest that the pattern exists in the Santa Cruz area.

The movement of sediment within the nearshore zone follows in a general way the cell-like circulation pattern. Both an onshore and an offshore movement of material of different sizes may occur at the same time; in addition there is a longshore movement of material by longshore currents, and a seaward movement by rip currents. Measured undertow velocities are low, generally less than 0.3 feet per second (Inman and Quinn, 1952, p. 24–36; Inman, 1954, p. 33), but rip currents have velocities up to 3 feet per second (Shepard, 1948, p. 43), and some authors consider these currents to be extremely important in the offshore transport of material (Shepard *et al.*, 1941, p. 364–365; Grant, 1943, p. 122–123). Rip currents are local features, but because they spread laterally outside the breakers, merging with other longshore currents, and because they may change position (Inman, 1953, p. 77; Trask and Johnson, 1955, p. 4), the deposition of sediment beyond the surf zone is not as localized as it might appear.

The fact that the modern platform is mantled by only a thin veneer of sediments indicates that its gradient is sufficient for the continuous removal of erosional debris, but this does not imply that the profile is one of equilibrium because the gradient could be steeper than the minimum one necessary for transportation. If submarine abrasion is restricted to very shallow water (Dietz and Menard, 1951, p. 1501; Fairbridge, 1952, p. 348, 352; Newell and Imbrie, 1955, p. 12; Bradley, 1956b), then the platform beyond the nearshore zone is not being modified today and hence its profile is probably not one of equilibrium. Equilibrium, if it exists at all, must occur only within the nearshore zone where it would relate to the size and amount of load moving in the complex cell-like pattern.

Summary of Characteristics

The modern wave-cut platform in the Santa Cruz area is an erosional feature at least 1½ miles wide. It slopes seaward at a low angle, generally less than 1°, and is slightly concave

[3] Mass transport is the forward drift of water particles moving in incomplete orbits.

upward within half a mile of shore where it coincides with the nearshore zone. The platform is either bare or mantled by less than 4 feet of sand whose grain size tends to be less than that of beach sand and appears to decrease slightly seaward.

The 100-foot Marine Terrace

General Statement

The lowest prominent marine terrace in the Santa Cruz area (Fig. 1; Pl. 1) contains an old wave-cut platform whose shoreline angle is now approximately 100 feet above sea level. Although warping has destroyed the original horizontality of the shoreline angle (Wilson, 1907, p. 14; Rode, 1930, p. 33; Alexander, 1953, p. 12–16), for convenience this terrace is here referred to as the 100-foot marine terrace; it is the same as Lawson's (1893, p. 140) 96-foot terrace, Wilson's (1907, p. 16) 112-foot terrace, the 100-foot terrace of Branner *et al.* (1909, p. 6), Rode's (1930, p. 33) and Alexander's (1953, p. 12) first terrace, and Page and Holmes' (1945) 90-foot terrace. The Monterey Formation cut by the old platform is similar in all respects to the Monterey cut by the modern platform. Occasional road cuts reveal the former sea cliff (Pl. 1; Fig. 2), buried by nonmarine detritus.

Old Wave-Cut Platform

The old platform of the 100-foot marine terrace at Santa Cruz, without its overlying terrace deposits, is shown in profile in Figure 6. The inner 800 feet of the platform are continuously exposed at the western city limits of Santa Cruz, and the rest of the profile is drawn through individually surveyed points shown by arrows in Figure 6. All irregularities of less than 5 feet were smoothed out in drawing the profile. Although terrace deposits prevent a continuous view of the whole platform, no evidence was seen to suggest that the 100-foot terrace might contain more than one former platform.

The old platform of the 100-foot terrace is similar in all respects to the modern platform (Figs. 5, 6). Its seaward inclination is concave within half a mile of the former sea cliff, where the gradient changes from about 1° to about 0.5° and becomes nearly uniform. Casual observation indicates that the platform has several feet of relief. The similarity between the two platforms suggests that both were developed in the same manner.

Terrace Deposits

Terrace deposits in the Santa Cruz area consist of a sheet of marine deposits overlying the old platform and a wedge of nonmarine deposits banked against the old sea cliff (Fig. 2).

Marine Deposits

Problem of genesis.—Different authors have suggested different modes of origin for marine-terrace deposits, and these suggestions have been grouped into six categories in Table 1. Some authors are listed in more than one category, because they interpreted parts of a specific deposit in more than one way. The grouping of authors as shown may not be entirely correct; some interpretations were not explicitly expressed and had to be inferred by the present writer, who accepts full responsibility for any misrepresentations that may have occurred.

With one exception the marine-terrace deposits in the Santa Cruz area have been called simply "marine." Rode (1930, p. 32–33) believed that the basal gravels of the deposits were left by a retreating sea, but that the overlying sandy material was continental in origin. Actually, the gravels constitute only a small portion of the marine deposits, for much of the overlying sand is also marine.

In view of the similarity between the old and modern wave-cut platforms, how do the marine-terrace deposits compare with the modern platform deposits? This author believes that the marine-terrace deposits are not the original neritic ones that existed when the platform was under water but instead are beach deposits which were prograded seaward during a period of emergence. The belief is based on the following evidence.

Thickness.—Marine-terrace deposits in the Santa Cruz area are commonly 5–15 feet thick and locally exceed 20 feet. Such thicknesses contrast with the thin sediments on the modern

TABLE 1.—INTERPRETATIONS OF MARINE-TERRACE DEPOSITS

TABLE 1.—INTERPRETATIONS OF MARINE-TERRACE DEPOSITS

Marine deposits, undifferentiated		Marine deposits, formed during submergence		Original neritic deposits*		Beach deposits, undifferentiated†		Beach deposits, formed during submergence		Beach deposits, formed during emergence	
Fairbanks	(1896)	Cooke	(1945)	Lawson	(1893)	Whitney	(1865)	Thompson	(1937)	Arnold and Anderson	(1907b)
Diller	(1901)	Krumbein	(1947)	Lawson	(1895)	Lawson	(1893)			Cotton	(1922)
Diller	(1902)	Alexander	(1953)	Diller	(1896)	Smith	(1898)			Rode	(1930)
Hershey	(1902)	California State Water Resources Board	(1953)	Diller	(1903)	Diller	(1899)			Grant and Gale	(1931)
Arnold	(1907)			Arnold and Anderson	(1907b)	Watts	(1900)			Capps	(1933)
Arnold and Anderson	(1907a)			Atwood	(1911)	Diller	(1902)			Davis	(1933)
Branner *et al.*	(1909)			Hanna	(1926)	Diller	(1903)			Anderson	(1936)
Lawson	(1914)			Smith and Mertie	(1930)	Tarr and Martin	(1912)			Twenhofel	(1943)
Weaver	(1916)			Smith	(1933)	Ellis and Lee	(1919)			Cotton	(1952)
Ellis and Lee	(1919)			Pardee	(1934)	Hanna	(1926)			Bradley	(This paper)
Hanna	(1926)			Webb	(1937)	Smith	(1933)				
Trask	(1926)			Cooke	(1945)	Pardee	(1934)				
Stephens	(1929)			Griggs	(1945)	Hertlein and Grant	(1944)				
Hoots	(1931)			Bruff	(1946)	Griggs	(1945)				
Anderson	(1936)			Wells *et al.*	(1946)	Bruff	(1946)				
Reiche	(1937)			Woodring *et al.*	(1946)	Wells *et al.*	(1946)				
Soper	(1938)			Cotton	(1952)	Woodring *et al.*	(1946)				
Flint	(1940)			Emerson and Addicott	(1953)	Washburn	(1947)				
Putnam	(1940)			Miller	(1953)	Hawkes	(1951)				
Putnam	(1942)					Upson	(1951a)				
Johnson	(1943)					Upson	(1951b)				
Allen and Baldwin	(1944)					Twenhofel	(1952)				
Hertlein and Grant	(1944)					Alexander	(1953)				
Baldwin	(1945)					Miller	(1953)				
Page and Holmes	(1945)					Toulmin	(1955)				
Woodring *et al.*	(1946)										
Flint	(1947)										
Beal	(1948)										
Upson	(1949)										
Weaver	(1949)										
Woodring and Bramlette	(1950)										
Upson	(1951a)										
Durrell	(1954)										
Hertlein and Grant	(1954)										
Jahns	(1954)										
Lian	(1954)										
Putnam	(1954)										
Toulmin	(1955)										

* Includes neritic deposits formed during emergence
† Includes beach deposits formed during stillstands of sea

platform but are comparable to modern beach deposits in the area.

Structure.—Laminations that contrast in grain size, sorting, and heavy-mineral content are characteristic of the terrace deposits (Pl. 2, fig. 1). Kuenen (1950, p. 287–288) stated that beach deposits are more variable in grain size than are neritic deposits, and laminations shown in Figure 1 of Plate 2 are considered by Thompson (1937), Martens (1939), Emery and Stevenson (1950), and McKee (1953) to be characteristic of beach deposits.

Black sands shown in Figure 1 of Plate 2 exist today in terrace deposits that are more than a mile seaward of the old sea cliff, as along West Cliff Drive in Santa Cruz, but they are absent from deposits closer to the old cliff. Black sands were undoubtedly present in all deposits at one time but are no longer recognizable near the old cliff because of leaching of unstable dark minerals. Even though weathering has partially destroyed black-sand laminae, laminations due solely to variations in grain size persist as recognizable features.

Cross-bedding is a common feature in the terrace deposits. Some cross laminations are developed on a large scale and dip 2°–8° seaward. The terrace deposits as a unit slope seaward at an angle of about 0.5°, and therefore the cross laminations within the deposits are discordant with the surfaces of the deposits. Large-scale gently dipping cross-bedding of this sort is considered a typical feature of beaches (Thompson, 1937; McKee, 1953). Some cross laminations are developed on a smaller scale and dip 25°–35° seaward or landward. Thompson (1937, p. 739–740) reported cross-bedding of this sort in lower foreshore deposits; McKee (1953, p. 5, 7, 9, 14) described similar cross-bedding in back-shore deposits. Because this type of cross-bedding was observed generally in the upper part of the terrace deposits, as would be expected of back-shore deposits, the writer favors McKee's interpretation.

Rock-boring mollusk holes.—The old platform is pitted with holes that were occupied by rock-boring mollusks when the platform was under water (Bradley, 1956a. Pl. 1), in the same way that the modern platform is pitted by the holes of living mollusks. Shells from some of the holes, exposed where erosion has stripped back the blanketing terrace deposits, have been dated older than 39,000 years (Bradley, 1956a). Because the terrace deposits are too thick for mollusks to have lived beneath, they must be younger than the mollusks; and because the mollusks lived in the erosional platform, the deposits must have been laid down after the platform was cut and therefore cannot be the original neritic deposits.

Grain size.—Unlike the modern platform deposits which tend to be finer-grained than beach deposits and appear to decrease in grain size seaward (Fig. 7), the terrace deposits everywhere contain material as coarse as beach deposits which shows no seaward decrease in grain size. Locally, even 1½ miles from the old sea cliff, the terrace deposits include marine gravels (Pl. 2, fig. 2). In addition to the coarse fraction, terrace deposits include a finer sand that resembles neritic material in Monterey Bay.

Mineralogy.—Terrace-deposit mineralogy is similar to that of modern beach and neritic deposits in the area, except for certain changes that are progressive within the terrace deposits. Although the terrace deposits are considered beach deposits, the mineralogy of neritic sediments is also discussed because evidence to be presented indicates that the terrace deposits are composed, in part, of reworked neritic material. The present discussion concerns only those aspects of the mineralogy that bear on the origin of the terrace deposits. All mineral percentages pertain to a specific size range and are used solely to show comparative variations between and within the various deposits.

Variations within the terrace deposits are revealed by the content of heavy minerals.

PLATE 1.—100-FOOT MARINE TERRACE

Near Laguna Creek

FIGURE 1.—100-foot terrace, looking south. Change in slope marks old sea cliff; higher terrace to the left.

FIGURE 2.—Old sea cliff of 100-foot terrace in road cut. Deposits at left are talus and slope-wash detritus. Shovel is 3 feet high.

FIGURE 3.—Nonmarine deposits of 100-foot terrace overlying marine with sharp contact at top of shovel handle. Shovel blade rests on old wave-cut platform. Shovel is 3 feet high.

FIGURE 1

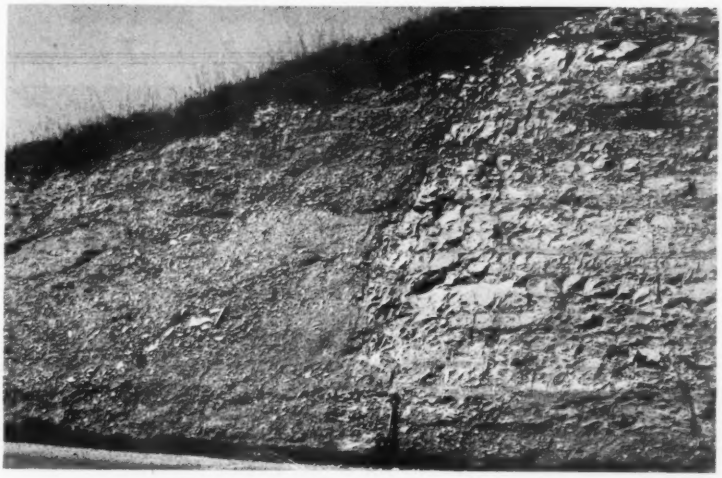

FIGURE 2

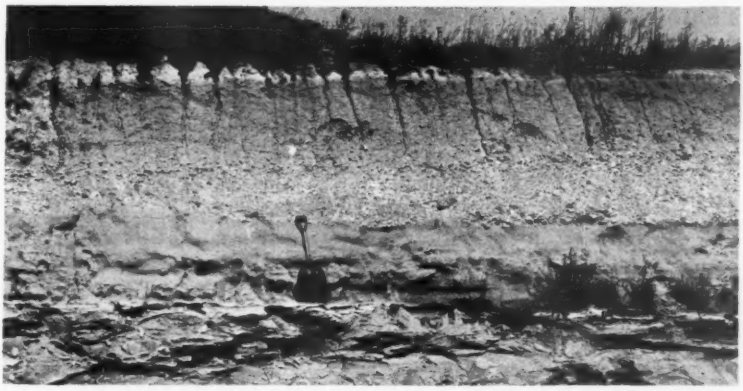

FIGURE 3

100-FOOT MARINE TERRACE

FIGURE 1

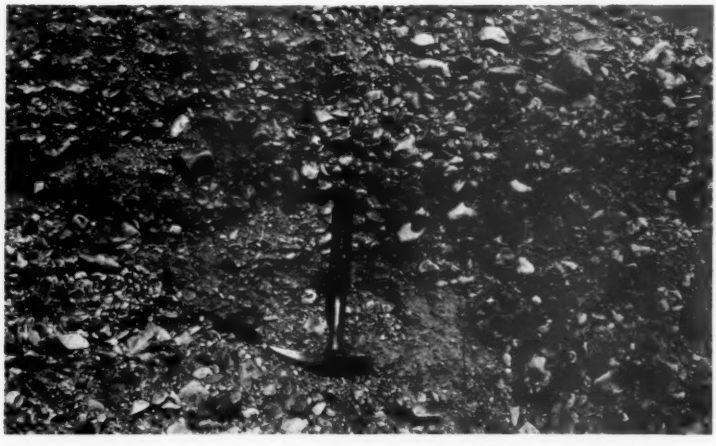

FIGURE 2

FIGURE 3

MARINE DEPOSITS OF 100-FOOT TERRACE

Figure 8 compares the heavy-mineral content of terrace deposits with that of beach and neritic deposits.[4] The heavies in the terrace deposits are less abundant than in the combined beach-neritic deposits and become progressively less abundant toward the old sea cliff; the average at *C* is 18 per cent, at *D* 12 per cent, and at *E* 7 per cent. If these deposits started out with approximately the same original content of heavy minerals, then the smaller quantity of heavies in the terrace deposits, compared with the combined beach-neritic deposits, suggests that the terrace deposits have been leached of some of their unstable minerals since deposition; the progressive decrease in the abundance of the heavies toward the old sea cliff suggests that leaching has been progressively greater in that direction. The average in column *C* would be higher were it not for the group of samples at the bottom which are believed to be reworked neritic deposits.

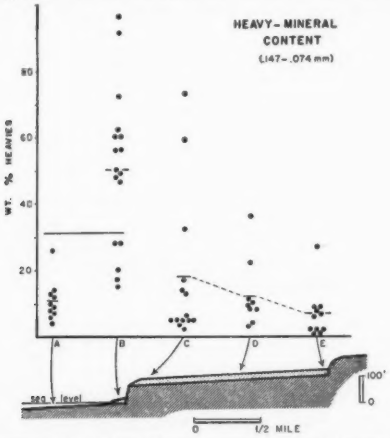

FIGURE 8.—COMPARISON OF HEAVY-MINERAL CONTENT BETWEEN NERITIC, BEACH, AND TERRACE DEPOSITS

Each dot represents one sample. Short solid lines indicate columnar averages; long solid line indicates combined beach-neritic average; dashed line indicates variation of columnar averages within terrace deposits.

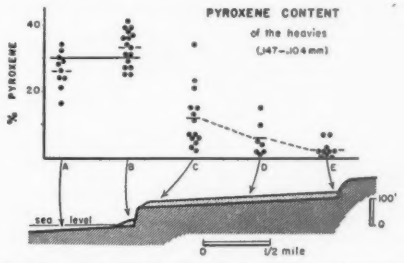

FIGURE 9.—COMPARISON OF PYROXENE CONTENT BETWEEN NERITIC, BEACH, AND TERRACE DEPOSITS

Each dot represents one sample. Short solid lines indicate columnar averages; long solid line indicates combined beach-neritic average; dashed line indicates variation of columnar averages within terrace deposits.

Variations in the content of heavy minerals are largely the result of variations in the content of pyroxene (Fig. 9). Pyroxene[5] is one of the most important heavy minerals in modern beach and neritic deposits, where the combined average is 30 per cent, but in the terrace deposits it is less abundant and it decreases progressively toward the old sea cliff; the average at *C* is 12 per cent, at *D* 6 per cent, and at *E* 2½ per cent. This decrease within the terrace deposits suggests that they have been leached of pyroxene to a progressively greater extent toward the old sea cliff.

Progressive leaching of pyroxene is shown also by the appearance of individual grains (Fig. 10). Pyroxene in modern beaches occurs in two general forms. One form, shown in

[4] In all mineral per cent diagrams, the terrace deposits are compared with the combined beach-neritic deposits. This is because the terrace deposits at *C* and *D* are composed in part of former neritic deposits which were reworked. It is not always possible to recognize reworked neritic deposits, and no attempt is made in columns *C* and *D* to distinguish the neritic constituents from the normal beach constituents. Therefore the averages shown in columns *C* and *D* represent averages of combined beach and reworked neritic sands and should be compared only with the combined modern beach and neritic sands.

[5] Augite, titanaugite, and hypersthene.

PLATE 2.—MARINE DEPOSITS OF 100-FOOT TERRACE

West Cliff Drive in Santa Cruz, 1½ miles from old sea cliff

FIGURE 1.—Laminated structure with black sand laminae.

FIGURE 2.—Marine gravels.

FIGURE 3.—Interlayering and mixing of coarse and fine marine sands. Fine sands are mica-rich.

sketch *A* of Figure 10, is euhedral to rounded and exhibits no etching. This variety of grain is found also in the Monterey Formation and was presumably derived therefrom. All pyroxenes within the terrace deposits have been

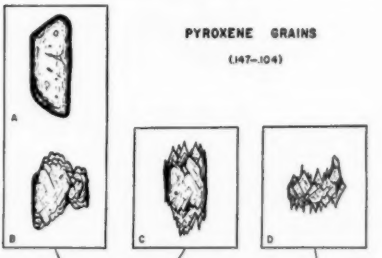

FIGURE 10.—COMPARISON OF APPEARANCE OF HYPERSTHENE GRAINS BETWEEN BEACH AND TERRACE DEPOSITS

Crystallographic *c*-axis is vertical in all grains.

etched by solution into sharp, cockscomblike forms shown in *C* and *D*.[6] Pyroxenes far from the old sea cliff *C* have been moderately etched and the original outline of the grains is still recognizable. Near the old cliff *D*, however, pyroxenes are so thoroughly etched that they have lost their original form. When the etched grains *C* fall onto the beach, their sharp points are rounded off in the surf zone, and they become grains as shown in *B*.[7]

In sphene there is a variation in content similar to that of pyroxene, but it is not so pronounced because sphene is less abundant (Fig. 11). In combined beach-neritic deposits sphene constitutes about 3 per cent of the heavies. The quantity of sphene in the terrace deposits is lower, and it decreases from 2 per cent at *C* to about 1 per cent at *E*.

[6] Similarly etched pyroxenes have been described by Ross *et al.* (1929), Edelman and Doeglas (1931), and Hutton (1952). The sand described by Hutton (p. 102–103) was probably a marine-terrace deposit similar to that discussed here.

[7] Hutton (1952, p. 73–74) observed that the blunting of etched pyroxenes takes place only on larger grains whereas those smaller than about 0.06 mm preserve their delicate etching even in beach environment. He attributed this preservation to the cushioning effect of a film of water around the finer particles. The writer's observations of pyroxenes of different sizes agree with those of Hutton.

Abraded sand-size shell fragments show a similar variation (Fig. 12). In the coarser sand fractions where they are most prevalent the fragments constitute 5–6 per cent of combined beach-neritic deposits. Shell fragments are

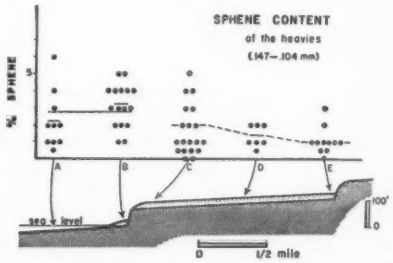

FIGURE 11.—COMPARISON OF SPHENE CONTENT BETWEEN NERITIC, BEACH, AND TERRACE DEPOSITS

Each dot represents one sample. Short solid lines indicate columnar averages; long solid line indicates combined beach-neritic average; dashed line indicates variation of columnar averages within terrace deposits.

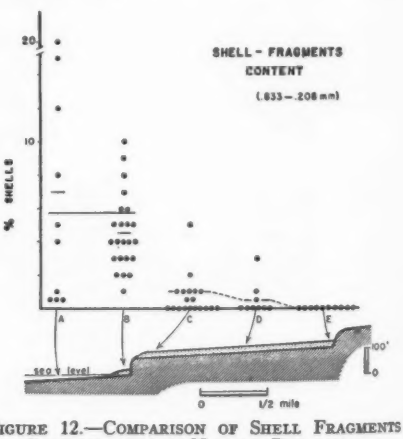

FIGURE 12.—COMPARISON OF SHELL FRAGMENTS CONTENT BETWEEN NERITIC, BEACH, AND TERRACE DEPOSITS

Each dot represents one sample. Short solid lines indicate columnar averages; long solid line indicates combined beach-neritic average; dashed line indicates variation of columnar averages within terrace deposits.

corroded and much less abundant in the terrace deposits; they constitute about 1 per cent of the sand at *C* and decrease until they are practically missing at *E*. The variation within the terrace deposits is probably due to progressive leaching, as with pyroxene and sphene. The lower content of shell fragments in the terrace

deposits, however, compared with the beach-neritic deposits, may also be due partly to the mechanical destruction of fragile shells during reworking of former neritic material.

Thus, leaching of terrace deposits has caused a progressive decrease toward the old sea cliff in the abundance of three relatively unstable constituents: pyroxene, sphene, and shell fragments. This variation suggests that the terrace deposits are not the same age everywhere, that they are oldest and most leached near the old cliff, and that they become progressively younger and less leached toward their distal portions. The leaching suggests also that emergence of the platform took place relatively slowly.

Summary of evidence favoring beach origin.—The above evidence suggests that the marine deposits of the 100-foot terrace are beach deposits prograded seaward during emergence. In thickness, in laminated structures that include black sands and cross-bedding, and in grain size, the terrace deposits resemble modern-beach deposits but contrast with neritic deposits. Burial of rock-boring mollusk holes and progressive decrease in age away from the old sea cliff are consistent with a theory of beach origin.

Genesis of Marine Deposits

General theory.—Figure 13 presents diagrammatically the writer's interpretation of the origin of the marine-terrace deposits. Sketch *A* represents the situation today. As sea level begins to fall (relatively), the beach is prograded seaward across the platform by the receding sea *B*. The source of sand for the expanding beach will be discussed shortly. If the process continues, the ultimate result will be an elevated wave-cut platform that is little altered from its original condition, mantled by a sheet of beach deposits *C*. The similarity between the old and modern platforms is evidence that the old platform was little altered during emergence. The interpretation of the Santa Cruz terrace deposits as prograded beach deposits agrees with similar interpretations by Arnold and Anderson (1907b) in the Santa Maria district of California, by Cotton (1922, 1952) in New Zealand, by Grant and Gale (1931), by Capps (1933) in Alaska, by Davis (1933) in the Santa Monica Mountains of California, by Anderson (1936) in Algeria, and by Twenhofel (1943) in Oregon.

Fine mica-rich sands.—The terrace deposits are composed of two general varieties of sand (Fig. 14). Group 1 consists of sands that are

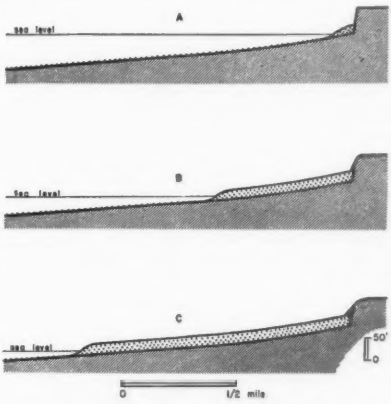

FIGURE 13.—FORMATION OF MARINE-TERRACE DEPOSITS DURING A PERIOD OF SLOW EMERGENCE

A is the present situation; *C* is the situation after emergence and progradation.

coarser than about 0.180 mm; their coarser fractions contain an average of 1 per cent or less mica. Group 2 consists of sands that are finer than 0.180 mm and relatively rich in mica; the mica is concentrated in the coarser fractions, as shown in Figure 3.

Modern beach deposits are similar to group 1 terrace deposits in that they are coarser than 0.180 mm and with one exception are poor in mica (Fig. 14). The sample containing a moderate amount of mica represents a special situation, discussed below.

Available neritic deposits are also plotted in Figure 14. The samples that are finer than 0.180 mm tend to be richer in mica than the others, and in this respect they resemble the group 2 terrace deposits. This resemblance is strengthened by Galliher's (1935, p. 1353) observation that fine mica-rich sands are one of the common bottom deposits in Monterey Bay. He found that there is little mica in sediments between the shore line and a depth of 30–60 feet, but that it becomes increasingly abundant in deeper sediments and is particularly concentrated at depths between 120 and 180 feet. The writer's neritic sample with the

most mica came from the deepest water (102 feet). This discovery seems to support Galliher's findings.

and the fine mica-rich terrace sands as former neritic deposits. The use of mica to recognize former neritic sediments was prophesied by

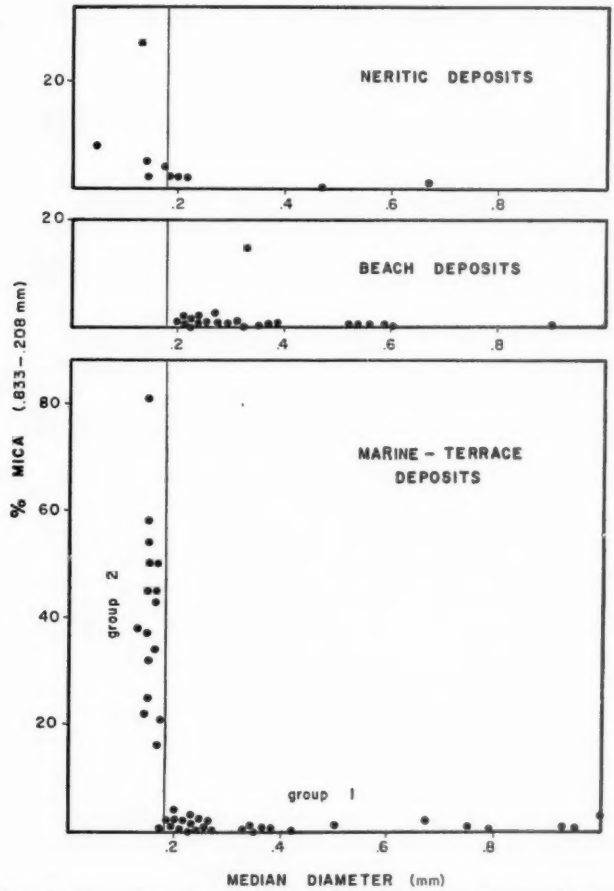

FIGURE 14.—COMPARISON OF GRAIN SIZE AND MICA CONTENT BETWEEN NERITIC, BEACH, AND TERRACE DEPOSITS
Line is drawn at 0.180 millimeters.

Group 2 terrace deposits further resemble modern neritic sediments in their content of heavy minerals. Neritic sediments contain a smaller proportion of heavy minerals than do beach deposits (Figs. 4, 8), and they compare favorably with the cluster of samples at the bottom of column *C* of Figure 8 which are almost exclusively fine mica-rich sands.

The coarse mica-poor terrace sands, therefore, are interpreted as former beach deposits, Galliher (1932, p. 57) when he said that biotite-rich sands " . . . might provide a reliable index for the interpretation of shallow deposits."

The areal distribution of fine mica-rich sands within the terrace deposits is consistent with the belief that they are former neritic sediments. In a zone about 0.3–0.5 miles wide next to the old sea cliff the terrace deposits consist entirely of the coarse mica-poor variety, and this zone corresponds to the zone of mica-

poor sediments adjacent to the present shore line. At a distance of 0.3–0.5 miles from the old cliff, however, fine mica-rich sands first appear in the terrace deposits, and they become increasingly abundant until at a distance of 1–1½ miles from the old cliff they locally constitute over half of the volume of terrace deposits.

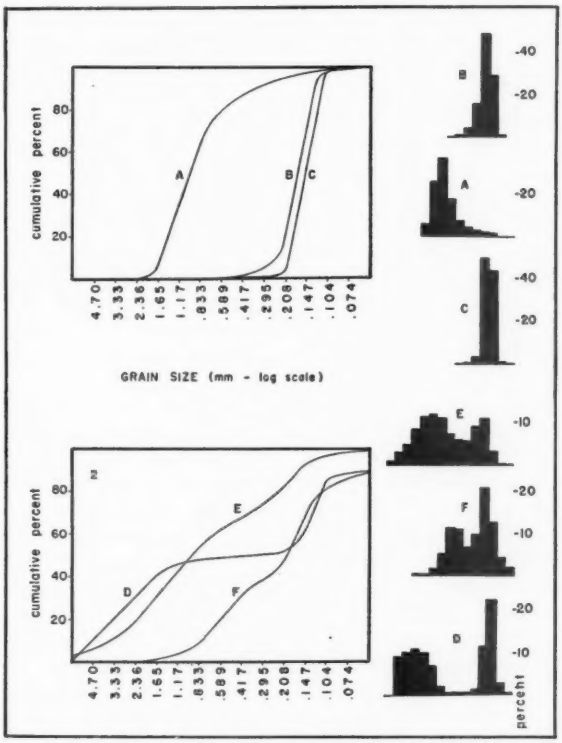

FIGURE 15.—CUMULATIVE CURVES AND HISTOGRAMS OF TERRACE DEPOSIT LAMINAE
Sketch *1* shows interlamination of coarse and fine sands; *2* shows mixing of the same types of sands. All fine samples are mica-rich.

Although the fine mica-rich terrace sands are considered former neritic sediments, two facts indicate that they are no longer in their original condition: (1) the thicknesses (over 8 feet in places) under which rock-boring mollusk holes are buried are too great; (2) they are interlaminated with, and in many places thoroughly mixed with, the coarse mica-poor sands. Figure 3 of Plate 2 shows the interlayering of the two types of sands, and Figure 15 compares histograms and cumulative curves of six terrace-deposit laminae. Sketch *1* of Figure 15 shows two fine laminae *B* and *C* separated by a coarser lamina *A*. All three are well sorted and unimodal, that is, there is only one peak in each histogram. The coarse lamina contains 1 per cent mica, the other two 45 per cent mica. Sketch *2* of Figure 15 shows three laminae in which the two types of sands are mixed (the coarse laminae in Figure 3 of Plate 2 contain curves of this type). These laminae are not as well sorted as those in sketch *1* and are distinctly bimodal, that is, their histograms show two peaks. An abundance of mica indicates that two separate sands are combined in each lamina: a coarser probably mica-poor variety and a finer mica-rich variety. Such bimodal curves suggest that sands from two

sources were deposited at the same time and were incompletely mixed (Pettijohn, 1949, p. 40; Inman, 1953, p. 76).

To summarize, the fine mica-rich terrace sands are believed to be former neritic deposits which were reworked into beach deposits during emergence. As sea level fell the prograding beach was constructed with whatever material was available (Fig. 13). The sea cliff, longshore drift, and streams supplied most of the beach material, and the offshore area made no permanent contribution to the beach because the long-term movement of debris was seaward. As sea level fell and the shore line receded, however, the sea cliff became a diminishing source of beach sand, but streams and longshore drift continued to be important contributors. Streams, extending across the new beach plain, added some of the unconsolidated deposits to their load from the hinterland and moved this augmented load out to the new shore line where it was redistributed by longshore drift. At the same time the offshore area became a new source of beach sand as the deep relatively quiet-water environment was transformed into an active shallow-water one, and fine mica-rich neritic deposits were then reworked into the prograding beach where they were interlaminated and mixed with the coarser normal-beach sands.

Mica, because of its shape, is winnowed from beaches and carried out to deeper quieter waters (Galliher, 1935, p. 1353; Inman, 1953, p. 37–40; 1954, p. 29; Wimberley, 1955, p. 34). Why was mica not winnowed from the neritic deposits while they were being reworked into the prograding beach? The writer believes that relatively little mica is being supplied to the beaches of the Santa Cruz area today. The Monterey Formation contains considerable mica, but less than 5 per cent of beach sand has been derived from the Monterey; this suggests that erosion of the sea cliff proceeds at such a slow rate that at any given locality the bulk of the beach sand has come from longshore drift rather than the immediate sea cliff. Rivers on Ben Lomond Mountain carry considerable mica, but only a relatively small amount of it reaches the beaches because the river mouths have recently been drowned, and most of the rivers today are separated from the ocean by bay-mouth bars. Only the three major rivers, the San Lorenzo, Scott Creek, and Waddell Creek (Fig. 1), maintain fairly continuous flow to the ocean; the San Lorenzo is the most important as a carrier of sediment. The one relatively mica-rich beach sample (Fig. 14) was collected at the mouth of the San Lorenzo, where there was an above-average supply of the mineral. The San Lorenzo made only a local contribution, however, for within a few hundred yards on either side of the river mouth the mica content of the beach was normal. Hutton (1952, p. 28, 30, 97) recorded a similar increase in the mica content of a beach at Pacific Grove, California, where the immediate sea cliff makes a significant contribution. Except locally, as at the mouth of the San Lorenzo River, the relatively small supply of mica is believed to be at least partly responsible for its scarcity in the beach deposits of the Santa Cruz area.

There is a large potential source of mica in the neritic deposits beyond the nearshore zone. These deposits do not contribute to the beach today because sediment movement is seaward, but they could become an active source if sea level fell. Consequently during emergence a large supply of mica, not now available, would be contributed to the beaches, and this could account in part for the greater abundance of mica on the prograding beach than on today's beaches.

The question still remains why the winnowing process did not keep the mica shifting seaward during the period of slow emergence that has been suggested. Theoretical considerations suggest an answer. Emergence of a concave sea floor whose gradient is one of transportation leaves the nearshore bottom with a slope that is too flat for the seaward movement of sediment (Fig. 13).[8] The original seaward movement may give way to an onshore movement in an attempt to restore the gradient by progradation of the beach. Wave activity at the shore should also be less at this time be-

[8] It is assumed that even though the load in the near-shore zone may decrease as wave activity is withdrawn from the sea cliff, the bottom profile is still too flat for effective dispersal of the load present. It is questionable whether the load would decrease very much; the sea cliff today is contributing sand at only a slow rate, and streams might have been more important in supplying material to the beach during emergence than they are today.

cause the broader zone of shallow water would dissipate much energy of the waves before they reach shore. Thus in time of emergence, in areas where the bottom slope was once one of transportation, an onshore movement of bottom sediments and reduced wave activity would favor the retention of mica on beaches.

The areal distribution of the fine mica-rich sands within the terrace deposits may be supporting evidence for the supposed onshore migration of neritic sediments during emergence. Galliher (1935, p. 1353) reported that Monterey Bay sediments richest in mica occur between 120 and 180 feet, and that sediments at depths less than 30–60 feet contain little mica. Some of the terrace deposits richest in mica, which possibly represent original neritic sediments from Galliher's 120- to 180-foot zone, now occur as close to the old sea cliff as 0.5 miles, where the depth would have been 40–50 feet. Even terrace sands that contain as much mica as the neritic sample from 102 feet (27 per cent) are now found 0.3 miles from the old cliff, where the depth would have been 30–40 feet. Although this situation may merely record an original inshore extension of the mica-rich neritic sediments, the occurrence of the same situation in three different marine-terrace levels suggests that during emergence the mica-rich sediments were moved inshore closer to the sea cliff.

Fine mica-rich sands are restricted to the lower part of terrace deposits and rarely contain black-sand laminae, whereas overlying deposits are coarse-grained and mica-poor and contain black sands. The lack of black sands in the fine mica-rich deposits is puzzling: even though neritic deposits contain relatively few heavy minerals, normally during reworking there would be a placering of available heavies into selected laminae. A possible explanation lies in the fact that where mica was observed on today's beaches, as at the mouth of the San Lorenzo River, it was restricted to the lower, flatter part of the beach.[9] These same deposits tend to have fewer heavy minerals (Fig. 16) and fewer black-sand laminae than do higher beach deposits, as has been reported by Pardee (1934, p. 20) and Griggs (1945, p. 146), presumably because concentration of heavy minerals is best accomplished near the limits of swash. It seems reasonable, therefore, that during reworking of neritic sediments, fine mica-rich sands collected on the lower flatter part of the beach where there was little tendency to concentrate the heavy minerals, and at the same time coarse mica-poor sands, brought in by longshore drift, collected on the steeper upper foreshore where they were winnowed into black and light laminae. As the beach was prograded seaward, upper foreshore deposits were extended over earlier lower foreshore deposits.

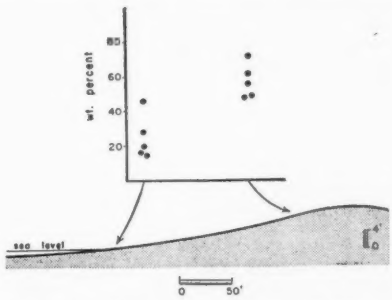

FIGURE 16.—COMPARISON OF HEAVY-MINERAL CONTENT BETWEEN HIGH AND LOW DEPOSITS ON BEACH

Black sand laminae are more conspicuous high on beach.

Genesis summary.—The marine-terrace deposits are beach deposits which were prograded seaward across the wave-cut platform during a period of slow emergence. They are composed in part of material similar to modern-beach deposits, supplied by streams and transported by longshore drift, and in part of former neritic material that was reworked during emergence. The former neritic material is still identifiable by its original characteristics: fine grain size, relatively high mica content, and relatively low heavy-mineral content. This material was redeposited on the lower part of the prograding beach at the same time that coarser sands, richer in heavies but poorer in mica, were deposited higher on the beach; mixing and interlamination of the two types of sands took place. Mica was retained on the prograding beach because of its abundance and because of reduced wave activity.

[9] Thompson (1937, p. 738–739) also reported a concentration of mica in lower foreshore deposits.

Nonmarine Deposits

Genesis.—Nonmarine deposits have accumulated near the old sea cliff ever since emergence began. They are a hybrid mixture of mass-movement and slope-wash detritus from the sea cliff (Pl. 1, fig. 2), alluvial-fan material, dune sand derived from the beach and banked against the cliff by onshore winds, and perhaps some beach sand thrown up during storms.

Distinguishing marine and nonmarine deposits.—Marine deposits are not always easily distinguished from nonmarine. There is little difficulty in distinguishing the two where the contact is sharp (Plate 1, Fig. 3), but in some places the contact is gradational, as has been reported in the Palos Verdes Hills of southern California (Woodring *et al.*, 1946, p. 55). Such contacts are most common near the base of the old sea cliff, where beach and dune sand was mixed with debris from the cliff.

Difficulty in distinguishing the two deposits arises also from their similarity in mineralogy: both contain a large percentage of quartz-feldspar sand, similar heavy minerals, and a few sand-size shell fragments. This similar mineralogy is due to the fact that nonmarine material was derived in part from underlying beach sand and in part from marine deposits of the next higher terrace exposed in the old sea cliff, and they therefore inherited some of the characteristics of the marine deposits.

Four criteria were helpful in distinguishing the two types of deposits: (1) Marine deposits are generally laminated, but nonmarine deposits are structureless or show only a crude stratification. (2) Less than 5 per cent of the marine deposits have been locally derived from the Monterey formation, whereas nonmarine deposits contain more locally derived material (Fig. 17). Where the nonmarine deposit is chiefly cliff talus (Pl. 1, fig. 2), Monterey material constitutes more than 80 per cent of the deposit. (3) Marine deposits contain little silt-clay matrix (less than 11 per cent), but the nonmarine contain noticeably more matrix (Fig. 17). (4) Marine deposits are usually well-sorted (sorting indices less than 1.5), whereas the nonmarine are less well sorted and commonly are very poorly sorted (Fig. 17).

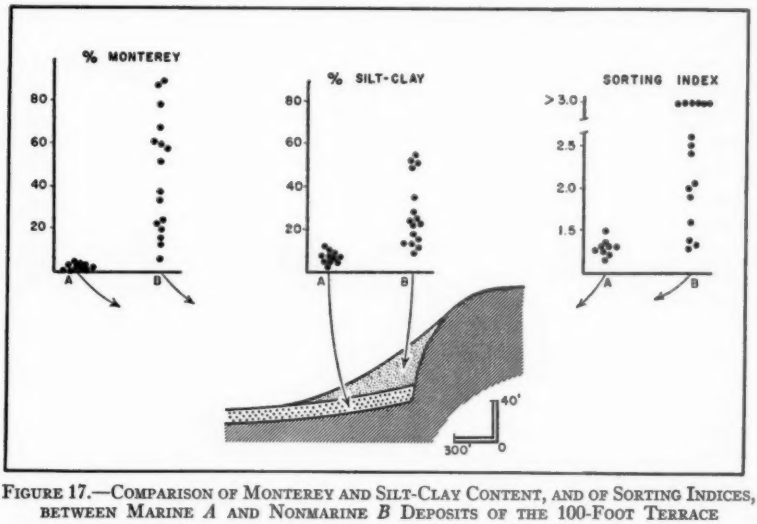

FIGURE 17.—COMPARISON OF MONTEREY AND SILT-CLAY CONTENT, AND OF SORTING INDICES, BETWEEN MARINE *A* AND NONMARINE *B* DEPOSITS OF THE 100-FOOT TERRACE

Figure 17 reveals that some marine and nonmarine deposits have a similar silt-clay content and similar sorting indices, a reflection of the hybrid nature of these nonmarine deposits. Nonmarine deposits derived principally from the sea cliff (Pl. 1, fig. 2) are high in Monterey fragments and a silt-clay matrix and have high sorting indices; those derived largely from

beach sand resemble the latter in their silt-clay content and sorting indices but contain slightly more Monterey material.

Higher Marine Terraces

Former Platforms

Platforms of higher terraces are similar to that of the 100-foot terrace, and they will not be described here.

Terrace Deposits

All higher unquestionable marine terraces contain marine deposits. These deposits are similar to those of the 100-foot terrace in (1) thickness and burial of rock-boring mollusk holes; (2) grain size, including the occurrence of fine mica-rich sands in two higher terraces; (3) laminated structures. Being older, these deposits are more weathered: more minerals have been leached, a greater content of silt-clay matrix has accumulated, and black sands, obscured by leaching even within some of the deposits of the 100-foot terrace, were never encountered.[10] Nevertheless, laminations due to variations in grain size alone have persisted as recognizable features in many places. Because of the similarity between higher and lower terrace deposits, a detailed discussion of the higher deposits is not included in this paper. The writer believes that the higher marine deposits are prograded beach deposits similar to those of the 100-foot terrace.

Weathered marine deposits of the higher terraces are easily confused in the field with some of the soft Miocene sandstones of the region. Branner *et al.*, (1909) and Rode (1930) called some terrace deposits north of Santa Cruz the Santa Margarita Formation. Laminated structures, topographic location, and mineralogy are the best criteria for recognizing higher marine deposits in the Santa Cruz area, and in some cases only a laboratory study will identify a sample.

High marine deposits are also difficult to distinguish from nonmarine deposits. Weathering has led to an increase in the silt-clay matrix and to a corresponding increase in the sorting indices of the marine deposits; consequently these two criteria, useful for differentiating the deposits of the 100-foot terrace, are not readily applicable to the higher deposits. Laminated structures and the content of locally derived fragments are the best criteria for distinguishing marine and nonmarine deposits of the higher terraces in the Santa Cruz area.

[10] Pyroxene, sphene, and shell fragments are rarely observed in higher terrace deposits, and hornblende is etched.

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MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, JUNE 15, 1956

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 68, PP. 445–468, 8 FIGS. APRIL 1957

VOLCANIC STRUCTURES IN THE CHUSKA MOUNTAINS, NAVAJO RESERVATION, ARIZONA–NEW MEXICO

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ABSTRACT

The Chuska Mountains, in northwestern New Mexico and adjacent Arizona, consist of flat-lying Miocene (?) sedimentary rocks locally overlain by pyroclastic and flow rocks of the Pliocene Navajo-Hopi volcanic province. The volcanic rocks were extruded into deep valleys cut into the top and flanks of the mountains.

Five volcanic centers within the Chuska Mountains were examined in detail. (1) The depression of Washington Pass is floored by three lava flows underlain by fluvial deposits of pyroclastic origin and overlain by a rubble dome intruded by plugs and dikes of minette. Minor caldera subsidence is indicated by faults in the Chuska sandstone underlying the volcanic rocks, by folds in the lava and tuff, and by steep inward dips of the volcanic rocks in the rim of the circular depression. (2) The Palisades represent an erosional remnant of a thick valley fill of coalesced domes of lava and agglomerate from several closely spaced vents. The flow rocks overlie fluvial beds of sandy tuff eroded from the slopes and deposited in the valley bottoms. The lava and tuff rest on slopes as high as 60°. (3) East Sonsela Butte is capped by three lava flows deposited in valleys cut into the Pliocene Chuska Mountains, and West Sonsela Butte is a remnant of a low, flat dome of lava and rubble capping a crater filled with coarse explosion breccia. (4) Tsailee Butte is a plug of columnar lava intruding a cone of coarse tuff-breccia, and South Tsailee Butte is a remnant of a fluvial crater filling of interbedded sand and tuff cut by a thick cone sheet of minette. (5) Roof Butte is formed from two funnel-shaped explosion pipes, one of which is capped by a low lava dome.

The vents of the Chuska Mountains have similar eruptive and structural histories. The volcanism was explosive in its early stages, and large quantities of pyroclastic material were ejected in numerous phreatic explosions. Between eruptions the craters of the low, flat cones were quickly filled with fluvial deposits of sand and tuff stripped from bordering valley walls. Collapse and inward sliding of the soft fluvial layers in the crater walls are shown by faults, crumpled and sheared layers, and folds. Volcanic activity ended with quiet extrusion of viscous lava and intrusion of dikes and plugs. The volcanic rocks consist of alkalic sanidine trachybasalt and minette.

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Introduction

The Chuska Mountains, the major highland of the Navajo Indian Reservation along the northern end of the Arizona-New Mexico State line, is an erosional remnant of flat-lying Tertiary sedimentary rocks intruded and overlain by scattered volcanic rocks of the Navajo-Hopi volcanic province. The range is 60 miles long and 10 miles broad and rises 2000 feet above the Defiance Plateau on the west and 4000 feet above the San Juan Basin on the east. The crest is a rolling upland at about 8500–9000 feet above sea level, and maximum local relief is 300 feet. The volcanic centers form the highest parts of the range: Roof Butte has an elevation of about 9600 feet, and the other large volcanic areas have roughly accordant summits at 9000–9200 feet.

About 36 necks and dikes occur in a zone 75 miles long in and near the Chuska Mountains. In six areas thick lava flows and pyroclastic deposits fill valleys cut into and through the Tertiary sedimentary rocks. Five of the areas were selected for detailed study to determine the nature of the volcanism and the relations of the volcanic rocks to the adjacent Tertiary sediments. The region has been described briefly by Gregory (1917), but the most pertinent earlier work is that of Williams (1936), whose study of the entire Navajo-Hopi volcanic province has now served as the basis for more detailed investigations in three portions of this large area—Hack (1942) on the diatremes, lavas, and sedimentary rocks of the Hopi Buttes; Balk (*in* Allen and Balk, 1954) on Buell Park and other volcanic centers in the Zilditloi field just southwest of the Chuska Mountains; and the present investigation of five centers in the Chuska Mountains.

The field work was done during parts of the summers of 1946 and 1950 with the aid of grants to Wright from The Geological Society of America and the Graduate School of the University of Minnesota. Field assistants in 1946 were H. T. Ames and J. H. Birman of Brown University, and in 1950 E. A. Hughes, K. E. Johnson, and A. V. Martini of Minnesota. The Navajo Indian Service allowed use of aerial photographs, quadrangle mosaics, and base maps. The writers are indebted to these individuals and organizations, to S. S. Goldich for petrographic advice, and to Clifford A. Kaye, Vincent Kelley, C. B. Hunt, and Howel Williams for criticism of the manuscript.

Regional Geologic Setting

The Chuska Mountains obliquely transect the eastward-dipping Defiance monocline, which separates the Defiance uplift from the San Juan basin. The maximum structural relief on the monocline is 5000 feet and at Hunters Point includes a fault with displacement of 2000 feet. Dips average 10°–15° E. but at Toadlena and Hunters Point are 90°. The monocline is domed at Todilto Park and near Beautiful Mountain.

Precambrian quartzite is exposed near Fort Definance and granite near Hunters Point. Permian and Mesozoic sedimentary rocks form hogbacks along the monocline, and earlier Paleozoic formations are recorded in the subsurface not far north and east (McKee, 1951, pl. 1). Most of these rocks are represented as xenoliths in the volcanic deposits.

The tilted Mesozoic rocks of the Defiance monocline are overlain unconformably in the Chuska Mountains by the flat-lying fluvial Deza Formation and eolian Chuska sandstone believed to be Miocene (?) (Wright, 1954, p. 1831; 1956, p. 427). The volcanic rocks of the Chuska Mountains, which are genetically related to the middle or late Pliocene rocks of the Hopi Buttes (Williams, 1936, p. 163), were extruded after the cementation and deep erosion of the Chuska sandstone.

Navajo-Hopi Volcanic Province

The Navajo-Hopi volcanic province is only one area of volcanism in a broad region characterized by great volcanic activity throughout the Cenozoic. It is distinguished by the relatively high proportion of pyroclastic deposits to lava flows and by the alkalic composition of the rocks.

The province comprises several separate volcanic clusters (Williams, 1936, Fig. 1), which may be grouped into 3 main areas (Fig. 1): the Hopi Buttes field (Hack, 1942), which includes more than 200 necks and diatremes and associated lava flows in a roughly circular area of 1500 square miles; the Monument Valley field, which is marked by about 20 necks, dikes, and diatremes; and, on the east aligned along the Defiance monocline, the scattered volcanic centers in and near the Chuska Mountains.

In the Chuska Mountains area (Figs. 1, 2) the southernmost intrusions are Twin Cones (Williams, 1936, p. 147) near Gallup and Black Butte south of Fort Defiance. Farther north is the Zilditloi field (Williams, 1936, p. 139; Allen and Balk, 1954, p. 99). Within the mountains are Washington Pass, most of the Wheatfields volcanic field, and the Lukachukai field. North of the Chuska Mountains and south of Carrizo Mountain is the Chuska Valley volcanic field, including Shiprock and near-by necks. The following five volcanic structures were selected for the present study (Fig. 2): Washington Pass; Palisades, Sonsela Buttes, and Tsailee Buttes in the Wheatfields field; and Roof Butte in the Lukachukai field.

Washington Pass

General Statement

The lava and tuff of the Washington Pass volcanic center bridge the narrow divide of the Chuska Mountains east of Crystal, New Mexico (Figs. 2, 3). Lavas occupy a circular depression 2 miles in diameter and 700 feet deep cut by sharp ravines at the eastern and western rims. The eastern entry is a narrow gap 100 feet wide bounded by sheer cliffs of tuff which rise 500 feet above the road. A cliff 600 feet high flanks the western portal.

Pre-Lava Surface

The pre-lava erosion surface cut into Chuska sandstone had moderate relief. Structural relations displayed in the high cliffs of tuff at the east and west sides of the depression show that the volcanic rocks were deposited on hill slopes and in valleys, but the depth of the valleys cannot be determined because of later subsidence.

The great thickness of fluvial sandy tuff in the cliffs south of Crystal Wash indicates that the material was probably deposited in a valley that sloped westward from a divide near the present top of the mountains. Subsidence of the area is suggested, however, by local folding and gravity faulting of the rocks beneath the volcanic cover. The lava extension south of Washington Pass shows no evidence of deformation, and its base defines the altitude of a wide depression in the top of the mountains, probably a valley whose stream drained northward to join the former westward-flowing stream near the present cliff at the western rim. A valley that extended eastward from the present eastern rim is indicated by thick deposits of fluvial sand and reworked tuff in the cliffs of the east rim and in the extension of tuff downslope from the rim. Here also, however, the pyroclastic deposits are faulted and sheared and show evidence of collapse.

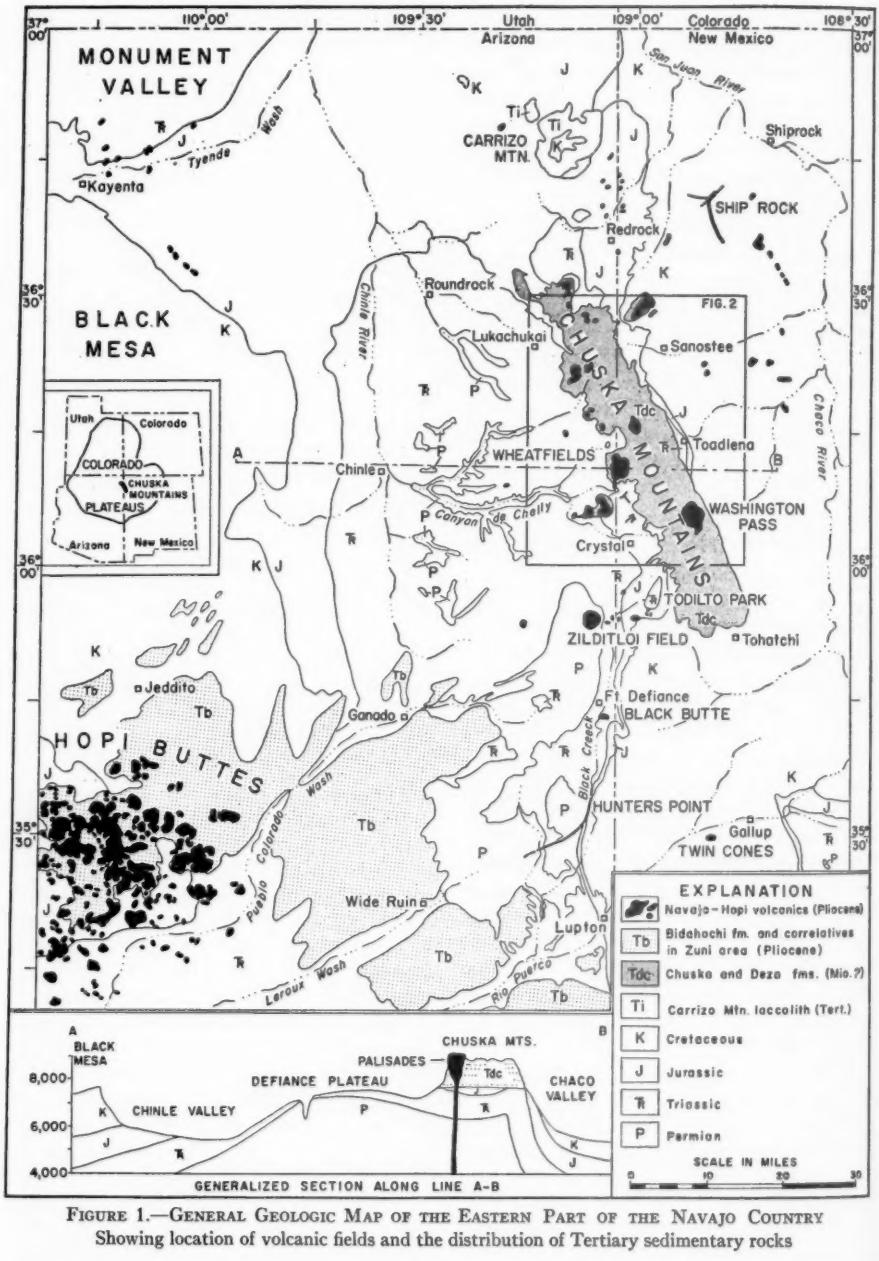

FIGURE 1.—GENERAL GEOLOGIC MAP OF THE EASTERN PART OF THE NAVAJO COUNTRY
Showing location of volcanic fields and the distribution of Tertiary sedimentary rocks

Pyroclastic Rocks

The first products of the volcanic activity were pyroclastic rocks, produced by explosive eruptions that characterize the opening of most Navajo-Hopi vents. The fragmental ejects are thickest along an eastward-treading line across the center of the circular depression. On the

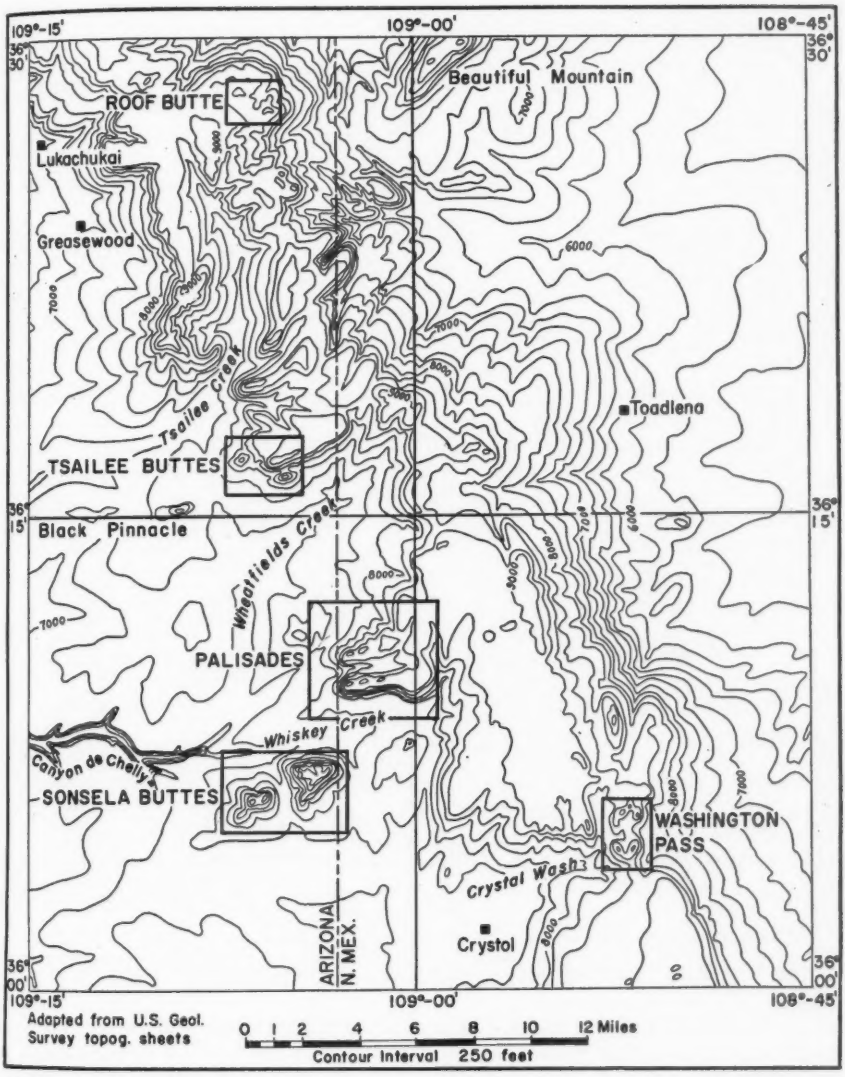

FIGURE 2.—TOPOGRAPHIC MAP OF ALL BUT THE SOUTHERN PART OF THE CHUSKA MOUNTAINS

eastern rim they attain a maximum thickness of more than 500 feet and thin markedly to the north and south. In the western rim they are 400 feet thick. On the northwestern and southern rims the pyroclastic beds beneath the lavas are less than 50 feet thick.

At the base the pyroclastic deposits are locally interbedded with sandstone that resembles the underlying Chuska sandstone but probably represents sand reworked therefrom. Elsewhere the pyroclastic deposits were laid down on the stripped surface of the cemented Chuska sandstone, and a sharp erosional unconformity separates the two formations.

The lower portion of the pyroclastic deposits consists mainly of buff sandy tuff and lapilli. Lenses of quartz sand containing scattered volcanic fragments are common. The high

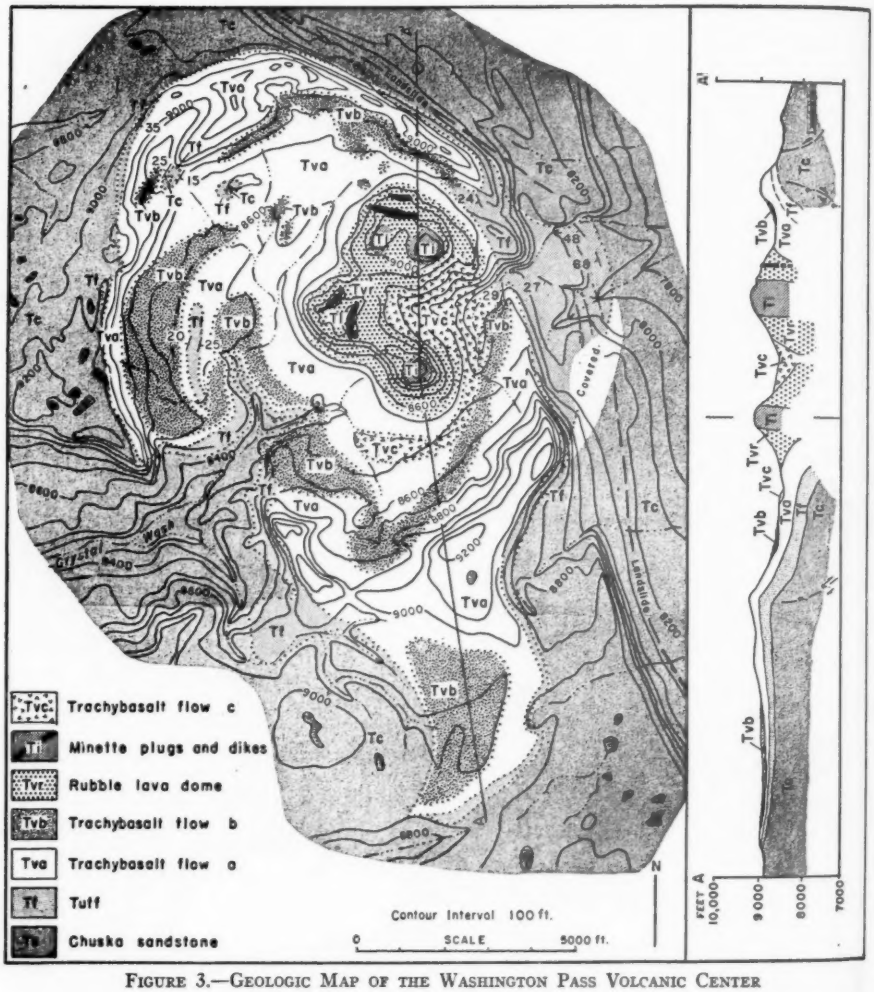

FIGURE 3.—GEOLOGIC MAP OF THE WASHINGTON PASS VOLCANIC CENTER

content of sand represents contamination from Chuska sandstone penetrated by the vent, plus wash from sandstone outcrops on the adjacent hill slopes. The sediments are generally well sorted and show graded bedding; grain size decreases upward in any unit. Minor disconformities, channeling, cross-bedding, and discontinuous and irregular beds indicate fluvial deposition. One bed of lacustrine limestone a few inches thick and bearing casts of plant stems was found south of the west portal.

Within the sequence of dominantly fluvial deposits and occurring with increasing frequency and thickness upward are beds 1–10 feet thick of poorly sorted nonbedded angular lapilli-tuff and breccia. Blocks of lava up to 12 inches in diameter are imbedded in the layers, and the depressions above them are filled by fluvial sediment. The beds of tuff beneath and surrounding the blocks were contorted by the impact of the falling blocks.

Overlying the predominantly water-laid beds of buff sandy tuff and lapilli with sharp angular unconformity are blue to blue-gray sandy lapilli-tuff and breccia. These beds have much the same characteristics as the underlying beds

but contain more coarse material. The change in color is probably due to the relative lack of quartz grains and the higher content of rock fragments, particularly dark lava.

Lava

The lavas of Washington Pass can be divided into three types on the basis of color and grain size, though all are similar mineralogically. They are all trachybasalt—extrusive equivalents of minette or biotite-vogesite. Lava fills the entire depression, covers most of the rim of Washington Pass, and extends southward 1 mile beyond the south rim. It attains a maximum thickness of 300 feet on the northeast rim.

The basal flow is a grayish-green fine-grained olivine-biotite trachybasalt. In the stream cut at the west rim it is less resistant than the overlying flow and weathers to a moderate slope broken by small rounded ledges. At the northeast rim, however, it forms a sheer cliff 300 feet high. The flow base is generally only slightly vesicular and contains rock fragments, tuff-breccia, and sand in a zone 3–6 inches thick.

The overlying rock is a dark-grayish-green to blackish-green lava darker and coarser-grained than the basal flow. The rock is well jointed and forms vertical columnar walls 20–50 feet high along the west rim. The joint planes extend through the contact into the basal flow, and the rock does not have the rubbly vesicular base found in many lava flows. Instead, the contact is commonly a relatively sharp, smooth break between the two rock types. There is only slight vesiculation and no noticeable lineation or platiness. This rock may therefore be a coarse-grained equivalent of the underlying lava and part of the same thick flow. It has been mapped separately on the basis of its dark color and coarse grain size.

The youngest lava in Washington Pass is a distinctive light-grayish-green to gray fine-grained porphyritic trachybasalt with small phenocrysts of olivine and biotite and vugs up to 2 inches in diameter filled with nests of olivine crystals. The rock is highly vesicular, locally almost pumiceous, and weathers to soft slopes and stream banks. Its maximum thickness is not known, as it lies in the semicircular depression east of the high peaks in the center of the pass. A narrow tongue extends around the south base of the peaks.

The high domelike area in the center of the pass is a mass of breccia, rubble, scoriaceous agglomerate, and dense lava intruded by three large columnar plugs and numerous dikes. The lava and scoria that make up most of the mass is trachybasalt similar to the lava that caps the rims and slopes of the pass. The rubble and agglomerate contain blocks of lava, tuff, and sandstone in a matrix of lava.

Intrusions

Three columnar plugs of dense black minette have intruded the dome in the center of the pass. On the south side of the northeast peak one plug is exposed in a vertical cliff 150 feet high, which is broken into large blocks by prominent cross joints. Inclusions of lava, tuff, and sandstone are more numerous toward the top. Near the top of the plug the rock becomes increasingly vesicular and at the top of the ridge forms a scoriaceous rubble that overlies and in part grades into the rubbly mass of the dome.

Intruding the rubble of the west peak are two dikes 20–70 feet wide. Vesicular cores in the dikes grade outward to dense glassy borders which have a marked vertical foliation parallel to the contacts. Rubble zones within the dike contain fragments of wall rock.

Some dikes peripheral to the central plugs on the northern flank of the northeast peak are similar to those of the west peak. In one dike a thin intrusion of almost pure feldspar (sanidine) abuts a dike of normal minette; the sanidine crystals continue unbroken through the contact. Small crystals of pyroxene, olivine, and biotite are scattered through the mass of feldspar.

Structure

The marked circular form of the depression of Washington Pass is most striking on aerial photographs. The pass is rimmed by a high ridge 9000–9200 feet in elevation and departs from a true circular form only by an offset of a quater of a mile in the west rim. The rim rock dips 25°–35° inward.

West of the central peaks and north of

Crystal Wash a system of arcuate folds deforms the pyroclastic rocks and lavas of the basin. These folds are peripheral to the peaks and roughly parallel to the circular rim.

beds inward and downward by amounts generally less than 1 foot. Two larger faults cutting the lower poorly bedded tuffs probably have displacements of as much as 7 feet. The

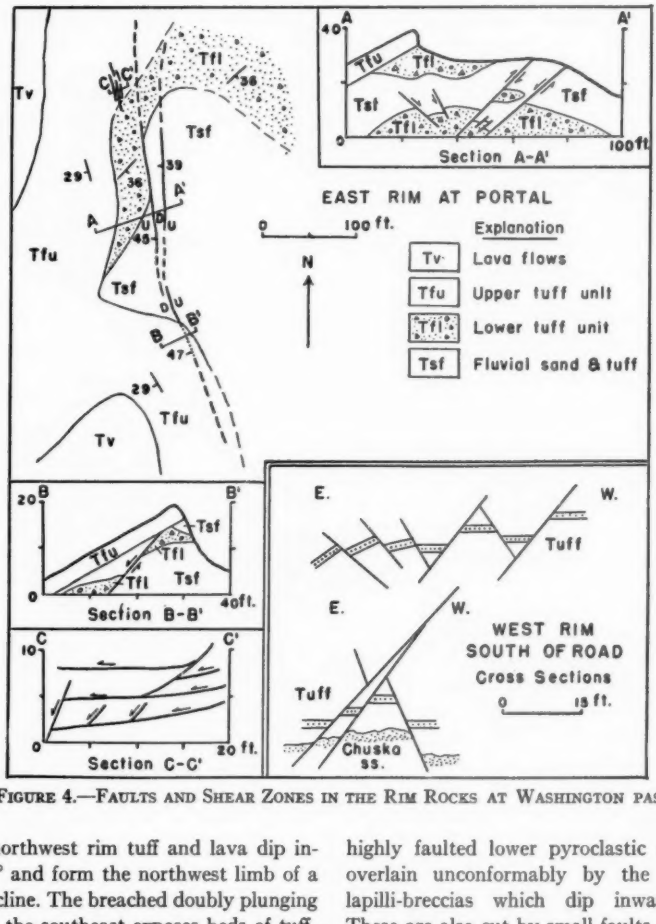

FIGURE 4.—FAULTS AND SHEAR ZONES IN THE RIM ROCKS AT WASHINGTON PASS

On the northwest rim tuff and lava dip inward at 35° and form the northwest limb of a narrow syncline. The breached doubly plunging anticline to the southeast exposes beds of tuff-breccia overlain by broken lava and underlain by Chuska sandstone in limbs dipping 25° N.W. and 15° SE. An anticline nearer the west entrance to the pass exposes tuff-breccia and lava in fractured limbs dipping 20° NW. and 25° SE. This fold is also exposed in the cliff bordering the north side of the road.

In the east rim the lower pyroclastic unit dips outward at inclinations of 5°–35° (Fig. 4). The beds are cut by discontinuous curved faults and fracture zones which displace the highly faulted lower pyroclastic deposits are overlain unconformably by the upper blue lapilli-breccias which dip inward 25°–30°. These are also cut by small faults and fracture zones.

In the west rim the outward-dipping tuff and sandstone of the cliff south of Crystal Wash also show systems of normal faults in an antithetic-synthetic pattern causing progressive downward and inward displacement of the beds (Fig. 4). Displacements are normally less than 2 feet but reach 5 feet in places.

The displacements of the pyroclastic deposits and interbedded fluvial sandstones are highly irregular; the fractures feather out into

contorted beds or disappear into unfractured rock. The irregularity and patterns of the faults indicate deformation of poorly consolidated material which yielded to deformational forces by expansion, compression, sliding, slumping, and folding in small increments without transmission of the breaks beyond short distances. The breaks became healed by compression packing during later adjustments and by cementation. Inward slumping was probably caused by crater subsidence.

The eastward extension of tuff and interbedded sand that floor the valley east of the east rim shows a much different structure. The unit is highly faulted, but the thick cover of brush, soil, and talus on the steep slopes renders it difficult to trace the structural elements. Beds of sandy tuff and lapilli breccia with thin interbedded lava, scoria, and agglomerate dip 25°-65° W. Three faults are exposed in the road cut below the east rim (Fig. 3). All dip 45°-50° E., and the beds are displaced downward to the east. These faults cannot be traced north or south because of heavy soil and brush mantle. The pyroclastic rocks in the ridge to the north are fractured, but no faulting could be demonstrated.

The deformation of this area may be the result of landsliding. The faults exposed in the road cut have the direction and inclination of landslide planes. North and south of the area landslides have occurred, and, as seen on aerial photographs, the valley and east rim of the pass are located between two of the large landslide blocks that cover the east flank of the Chuska Mountains.

Sequence of Events

Before volcanism, erosion by a westward-flowing stream had produced a shallow valley; a tributary valley entered from the south. An eastward-flowing stream may have cut a valley near the present east portal.

The first volcanic activity, the explosive eruption of fine ash and lapilli, probably came from a conduit in the shallow pass between the two main valleys. The explosions probably resulted in a low pyroclastic cone around the vent and deposited a thin layer of ash and lapilli on the surrounding slopes. Erosion of the cone and the hills was accompanied by deposition of thick fluvial sandy pyroclastic materials in the valleys; numerous layers of poorly sorted breccias and short lava flows indicate renewed volcanic activity. At least one small lake was formed in a depression or valley choked with debris, and in it was deposited a thin bed of limestone.

The main vent was probably near the present east rim, for here are located the three plugs that later intruded the crater. Here also are certain features not on the west rim, such as several thin flows of lava and agglomerate and also unconformities within the pyroclastic deposits.

Deformation accompanied and followed the explosive eruptions. It was probably caused by enlargement of the crater by inward slumping of the crater walls induced by explosions and loading in soft, unconsolidated pyroclastic materials. The funnel shape of explosion pipes and the related disruption of crater walls have been demonstrated in a number of localities (Wagner, 1922, p. 27; Hack, 1942, p. 367) and are found in other vents of the Navajo-Hopi region. Though the actual crater walls are covered by lava in Washington Pass, the early activity recorded by the explosive products indicates a history and crater evolution similar to that of other vents.

After this episode of deformation came the relatively quiet extrusion of lava. The fine- and coarse-grained trachybasalts may represent two flows or variations of one thick viscous flow. Near the vents north of the peaks the flows are more than 300 feet thick. At the west rim a thickness of more than 80 feet is preserved in the valley of Crystal Wash. Probably the lava formed a low dome around the vents, flowed outward and filled the depressions around the crater. Lava flowed more than 2 miles south from the vent area. Blocks of lava on the knoll at 9200 feet a quarter of a mile west of the west rim may represent a western extension of the flow. The northeastern and southeastern rims are steep landslide scarps on the flank of the Chuska Mountains, and the extension of lava in those directions cannot be determined.

After the extrusion of the large lava flows crater subsidence estimated at 300 feet occurred. This second episode of subsidence is indicated by steep inward dips of the lavas and

pyroclastic deposits and by arcuate folds in the basin floor.

The usual mechanism to account for caldera subsidence is withdrawal of magma to below the top of the magma chamber. Through caving of the overlying material and transmission of ring fractures to the surface, a circular block is dropped, and a depression or caldera is formed at the surface (Williams, 1941, p. 333). In competent lava or other hard rock this caldera is ringed by a fault, but unconsolidated materials tend to slump and slide inward in a funnel-shaped section with decreasing deformation outward and upward from the point of initial slumping. Cloos (*in* Balk, 1938, p. 30) has demonstrated this process in photographs taken of slumps in wet clay. Williams (1936, p. 120) calls upon essentially the same mechanism for the caldera subsidence at Indian Wells in the Hopi Buttes area.

The Washington Pass subsidence is believed to have been initiated in the conduit beneath the surface. Explosive eruptions and extrusion of lava increased the size of the crater and gradually enlarged the conduit, particularly in less resistant rocks like the Chuska sandstone. Gradual stoping of the walls through collapse and assimilation during stillstands of the lava possibly also enlarged the conduit.

The subsequent extrusion of a relatively large quantity of lava in the form of a dome concentrated a tremendous weight on this weakened, fractured zone of Chuska sandstone. Progressive inward slumping after withdrawal of the magma from the enlarged conduit then initiated caving. Caving ceased when the conduit was filled and the weight of the overlying dome was supported.

The final phase of volcanic activity was the extrusion of the rubble dome which today stands in the center of the pass. The dome has been intruded by plugs and dikes of minette, forming the three peaks of the dome. A small lava flow, extruded possibly from a dike in the center of the dome, fills the trough between the dome and the south-rim slope of the basin.

Erosion and landslides since volcanism have cut gaps into the caldera and have removed the upper parts of the valleys in which the volcanic rocks were deposited. The rubble and agglomerate of the sides of the central dome have been removed, and a steep-sided remnant in which faint horizontal banding shows its exogenous origin remains.

The Palisades

General Statement

The Palisades form the imposing western rim of the Chuska Mountains. They rise to an elevation of 9200 feet above sea level, 1800 feet above the lowlands north of Whiskey Creek (Fig. 2). They are aptly named, for the south wall, more than a mile long, is a vertical cliff of massive columnar basalt over 400 feet high. The mass is a roughly triangular pile of lava, breccia, and tuff deposited in a large valley system cut into Chuska sandstone (Fig. 5).

Pre-Lava Surface

The erosion surface on which the volcanic rocks were deposited had considerable relief. At the southwest corner and on the north side, lava and tuff rest on old hill slopes cut in Tertiary and Mesozoic rocks at elevations ranging from 7600 to 8400 feet. The lava on the eastern rim was deposited on a valley slope cut into Chuska sandstone at an elevation of 8800 feet. Thus there was relief of at least 1200 feet at the time of volcanism.

The large volcanic mass of the Palisades was deposited in a broad valley system that drained into the Pliocene Wheatfields Creek. The southernmost lava remnant of the mass (7900 feet) and the northernmost (7700 feet) mark the extensions of the main valley which had a width of about a mile and an over-all gradient of about 50 feet per mile. Smaller, more steeply sloping tributary valleys branching eastward and westward from the larger valley are outlined by extensions of lava beyond the cliff of the Palisades. A lava remnant at the northwest corner fills a valley that was probably a northward-draining tributary of the main valley. At the southwest corner a steeply sloped lava remnant covered the wall of a valley that drained westward into the Pliocene counterpart of Whiskey Creek.

The topography of the mountain front at the time of volcanism was similar to that of today; steep valley walls were eroded into a succession of scarps and slopes formed in alternating layers

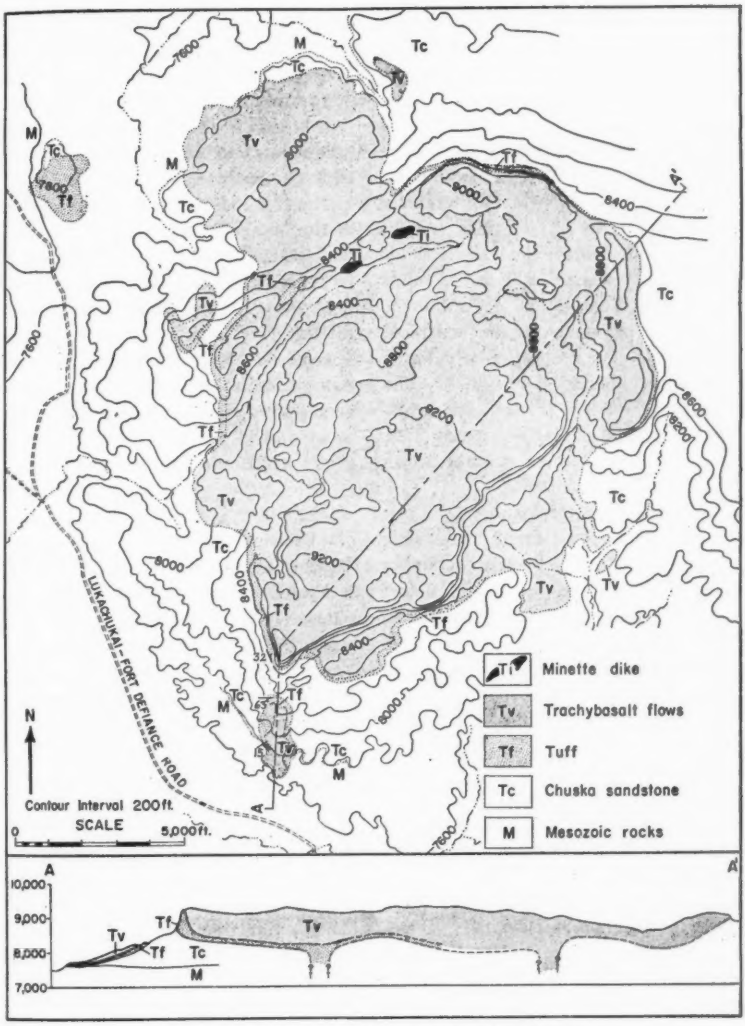

FIGURE 5.—GEOLOGIC MAP OF THE PALISADES

of well-cemented and poorly cemented Chuska sandstone. On the old hill slopes the contact between the volcanic rocks and the underlying sandstone is sharply erosional and inclines about 30°. Low escarpments, inclined as much as 60°, break the gentler slopes. In the bottoms of the valleys the volcanic rocks are interbedded with fluvial sands on gradients of 5°–10° and cover low ridges which have side slopes as high as 30°.

Pyroclastic Rocks

At most localities in the Palisades, tuff and tuff-breccia overlie Chuska sandstone and underlie the lava. Locally, as at the northeast corner, flow breccia and lava rest directly on sandstone. The southern and northern extensions of the lava beyond the sheer wall of the Palisades are concealed by talus and heavy forest cover.

The southwest corner of the Palisades and the neighboring detached lava remnant offer the best section. The volcanic rocks are unconformably underlain by white to buff cross-bedded Chuska sandstone. The contact is sharply erosional and represents a valley slope inclined about 30° and broken locally by escarpments. Fluvial deposits of gravelly tuffaceous sandstone 10–50 feet thick rest on old valley floors cut through the Chuska sandstone into Triassic red shale. The fluvial sands are faintly bedded in lenses and discontinuous cross-bedded layers and contain scattered volcanic fragments. The proportion of volcanic material increases upward, and the deposits pass gradually to crudely layered and cross-bedded white to light-olive-green sandy tuff, lapilli, and tuff-breccia which contain layers and lenses of silicified sandstone.

Overlying the sandy gradational units in this area are bedded, poorly sorted pyroclastic deposits ranging from tuff to breccia containing huge blocks. The point of one block 10 feet wide by 15 feet high is imbedded in sandstone beneath the pyroclastic layers, and the surrounding sediments and tuff are contorted and faulted. Irregular layers of sandstone and sandy tuff within the pyroclastic bids are locally silicified, colored bright lavender to red, and stained black near the contacts.

Small shear zones and echelon faults in the pyroclastic rocks are oriented in three directions at 50° and at 45° to the bedding planes, and show irregular displacement of 1–6 inches. The Chuska sandstone underlying the pyroclastic rocks is sheared and faulted; it was probably indurated at the time of deformation. Well-cemented layers are broken by fractures that feather into undisplaced poorly cemented layers. Most of the pyroclastic rocks show crumpled bedding planes, discontinuous healed faults, and other indications that the deposits were not consolidated. Some cracks filled with calcite are more resistant to weathering than are the surrounding sedimentary rocks. Extreme shearing and crumpling is most noticeable on steeper slopes, apparently because of load and slumping of unconsolidated slope debris and poorly cemented units of Chuska sandstone. Deformation is less in more highy cemented sandstone units and on gentler slopes.

At two places on the south wall of the Palisades the flows were extruded against nearly vertical escarpments of Chuska sandstone. Here the wall of lava is veneered with a coarse flow-breccia of lava, sandstone, and tuff which grades into rubble and tuff-breccia where the contact is less steep.

The tuff and breccia of the Palisades contain in the lower 50 feet principally coarse-grained dark-green olivine-biotite trachybasalt, and in the higher units mostly fine-grained gray to light-green olivine-biotite trachybasalt similar to the flow rock that caps the Palisades. Sparsely scattered throughout the section are fragments of basic porphyritic lavas. Small xenoliths of granite, syenite, and metamorphic rocks and of red and white (Mesozoic-Paleozoic?) sandstone are numerous.

Lava

The mass of flow rock that forms the sheer lava wall of the Palisades is predominantly light-olive-green to light-blue-gray porphyritic trachybasalt, containing numerous phenocrysts of olivine up to a quarter of an inch in diameter and scattered phenocrysts of augite, sanidine, and biotite. With increase in grain size the color of the rock changes to dark green.

Spectacular vertical lava columns as much as 400 feet high form the south wall of the Palisades. The columns, 5–10 feet in diameter, present a striking "organ pipe" appearance where the edges have been rounded by weathering. Elsewhere the jointing is variable and in random directions. At places the flows weather in nodes or spheroids, but distinct platiness is developed at other localities. Interbedded layers of agglomerate, breccia, and scoria occur locally, but the mass is largely dense to vesicular lava.

At several localities on the Palisades, low cones of scoriaceous lava and agglomerate suggest small conduits. Scattered inclusions of tuff, breccia, and lava in a matrix of fine-grained blue-gray scoria form rubble mounds that locally grade into the surrounding lavas.

Intrusions

Near the north rim a coarse-grained black to blackish-green dense minette intrudes the fine-

grained lavas. Contacts with the lava range from horizontal to vertical, and the surrounding lava is baked red up to an inch from the contact. The intrusion is exposed in two small patches and may be a dike.

A volcanic neck is located on the northward trending ridge 1 mile north of the northwest corner of the Palisades. The vent agglomerate is a comminuted mass of tuff and lapilli breccia, blocks of lava, and many xenoliths of acidic igneous rock and red shaly sandstone in a matrix of scoriaceous to dense coarse-grained dark-green prophyritic minette containing phenocrysts of olivine, augite, sanidine, and biotite. A section 100 feet thick of poorly sorted crudely bedded lapilli breccia is cut by apophyses of the minette.

Sequence of Events

Volcanism commenced with explosive eruptions, probably from several closely spaced pipes within a large valley, and added tuff and lapilli to the fluvial sands. Explosions of increasing intensity are recorded by coarser volcanic debris higher in the section. That lava occasionally plugged the craters is shown by the large blocks of flow rock in the breccias. Erosion of the mountain front continued throughout this period and contributed sand to the ash and breccia falls. Cross-bedding, pebble and sand interbeds, and local ripple marks indicate stream deposition during intervals of quiescence.

The lavas of the Palisades were extruded in large quantities relative to those in other areas of the Chuska Mountains. Low domes near the vents enlarged and coalesced into a thick sheet which eventually filled the valley and spilled over the lowest divides into adjoining depressions. There is no evidence that the lavas extended much farther than the limit of the remnants shown on the map. The small vent at the northwest corner of the Palisades is composed of tuff-breccia and does not show evidence of having extruded large quantities of lava. A neck and associated tri-radiate dikes south of Whiskey Creek may have contributed lava both to the Palisades and to Sonsela Buttes.

Sonsela Buttes

General Statement

Sonsela Buttes rise to 9000 feet above sea level at the head of Canyon de Chelly and form an imposing outlier of the Chuska Mountains (Fig. 2). East Sonsela Butte, the higher and larger remnant, is capped by a series of flat-lying lava flows, West Sonsela by a thick dome of lava and tuff (Fig. 6).

The erosion surface on which the volcanic rocks were deposited was a system of ridges and valleys cut in the formerly more extensive mountains, decreasing in elevation from 8600 feet at the eastern side to 8200 feet at the western. Local relief amounted to only about 200 feet, but the mountain front dropped abruptly to a lower level west of the Sonsela Buttes, where a lava flow caps a surface cut on Triassic shale at 7600 feet.

East Sonsela Butte

East Sonsela Butte is capped by at least three flows which rest in most places directly on Chuska sandstone. Tuff occurs locally in thin discontinuous layers at the base of the volcanic sequence. Where flows overlie Chuska sandstone with no intervening pyroclastic deposits, the sandstone is locally baked and hardened and contains irregular buff to red silicified zones and well-developed shaly partings parallel to the contact. Rubble at the base of the lava is rare, although tongues of vesicular lava and interlayered breccia and agglomerate are found locally. At most places the contact is regular, as if fluid lava flowed smoothly downslope.

The first lava flow, a dark-green medium-grained trachybasalt, caps the eastern and northern parts of the butte and ranges in thickness from 20 to 60 feet. It decreases in elevation and thickness from east to west and may have been extruded from a source farther to the east, perhaps from the cone and associated radial dikes 2 miles east of Sonsela Buttes.

The second flow, a light-gray to blue-gray fine-grained trachybasalt, forms low hills in the central and southern parts of the butte. It is 50–90 feet thick in its easternmost extensions but thickens rapidly westward and forms sheer cliffs a maximum of 200 feet high in the western

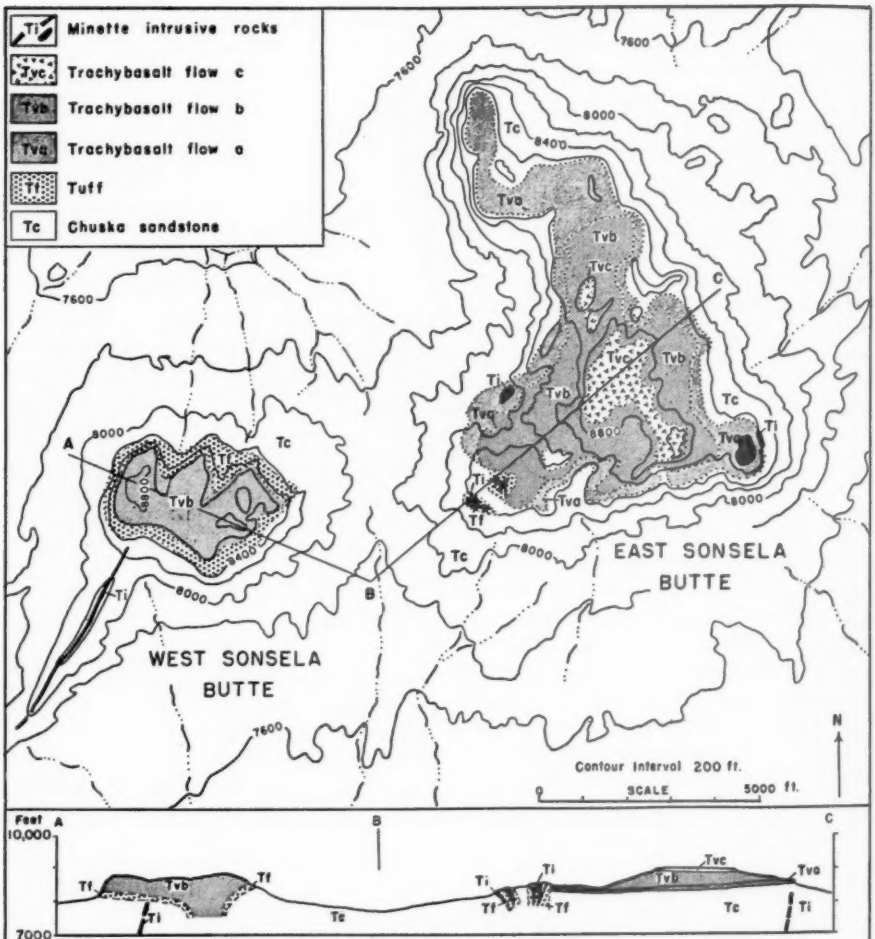

FIGURE 6.—GEOLOGIC MAP OF SONSELA BUTTES

rim. It is very similar to the lava of West Sonsela Butte.

The third flow is a relatively thin medium-grained olive-green porphyritic trachybasalt. It has a thickness of about 50 feet and caps the highest point of the butte.

East Sonsela Vent

A small volcanic neck in the pass between East and West Sonsela Buttes may mark the source of part of the lavas of the buttes. About 30 feet of tuff-breccia is exposed in a vertical neck intruding Chuska sandstone. Apophyses of minette and blocks of breccia, lava, sandstone, and granite form a vent agglomerate intruding the tuff-breccia. Lava of the second flow of East Sonsela Butte overlaps the east side of the vent area.

Farther downslope to the southwest, lava and tuff outcrop in a talus-covered slope of sandstone. Lava and coarse tuff-breccia form a thick, curved dikelike body dipping 60°–90° E. Tuffaceous sandstone overlies the igneous rock; Chuska sandstone underlies it. The area may be a collapse zone in which the crater

filling has slumped into a large conduit, or it may be an intrusive on the flank of a larger cone.

West Sonsela Butte

West Sonsela Butte is a circular, dome-shaped mass of lava capping a pyroclastic-filled crater about a mile in diameter. Outcrops on the north side of the butte are concealed under heavy talus and forest cover, but elsewhere the domical structure of the lava cap is revealed by crude horizontal bedding in near-vertical cliffs as much as 200 feet high.

The pyroclastic deposits underlying the lavas are poorly sorted brown to light-green tuff, lapilli-tuff, and breccia, and have a maximum thickness of about 50 feet. Fragments of volcanic rock, sandstone, and biotite granite are numerous in the breccia. The basal tuff contains lenses of sand and is crudely cross-bedded as if reworked by streams. The higher units are nonbedded; they contain a large proportion of coarse fragmental lava and breccia and are interlayered with tongues of dark-green platy lava and agglomerate.

The lava cap of West Sonsela Butte is predominantly gray fine-grained trachybasalt similar to the second flow of East Sonsela Butte. Interbedded with it are tongues of agglomerate and dark-green lava. The west side of the butte is a cliff more than 100 feet high of crudely banded lava and agglomerate. The eastern cliff reveals a thick section of lava, agglomerate, and rubble cut by dikes and stringers of black dense minette.

The dome was apparently formed by extrusion of lava from a central crater. Flows, welling out in thick viscous tongues, pushed and over-rode the earlier lavas and pyroclastic deposits and built a low rubble dome around the vent. The central conduit has been revealed by a notch eroded in the eastern face of the butte. There a vertical funnel-shaped mass of rubble intrudes a massive scoriaceous agglomerate that is continuous with the lava and agglomerate forming the dome.

There is no evidence of explosive eruption and rupturing of the dome characteristic of viscous rapidly formed domes in large volcanoes (Williams, 1932, p. 141). Instead, the alkalic lavas of East Sonsela Butte seem to have welled out quietly following the initial explosive activity, in a manner similar to that of the Hopi vents described by Hack (1942, p. 363). Stearns (1926, p. 356) describes large exogenous basaltic domes in which the summit craters are still preserved. Erosion of Sonsela Buttes has removed all but a small part of the lava.

The volcano of West Sonsela Butte was probably built in a wide, gently sloping depression near the edge of the mountains. The lava, after overflowing the low pyroclastic cone which had been built around the vent, filled the valley to the east, at the central and western part of the present East Sonsela Butte, and formed the thick flow which caps that butte.

Sonsela Dikes

The lavas and tuffs of Sonsela Buttes are intruded by dikes composed of black to dark-green fine-grained slightly porophyritic minette, very similar to the rocks forming the plugs at Washington Pass and the dike at the Palisades.

On East Sonsela Butte the intrusion is found in the vent area and in a small patch northeast of the vent cutting the fine-grained flows. It is also found on the southeast rim underneath, but not intruding, the lava and enclosing a block of sandstone stoped from the wall. The sandstone inclusion and wallrock are silicified and baked reddish brown.

On West Sonsela Butte the rock intrudes the agglomerate and lava of the east face and forms apophyses and dikes in the vent rubble. Extending southwestward from the butte is a dike 10–50 feet wide and almost 2 miles long. Contacts with the Chuska sandstone are vertical and show slight heat effect. Locally the borders of the dike are rubbly and vesicular, and small stringers extend into fractures in the hardened wallrock.

Tsailee Buttes

General Statement

Three miles north of Wheatfields Lake, near the headwaters of Wheatfields Creek, are two prominent buttes rising about 1800 feet above the bordering lowlands (Figs. 2, 7). Tsailee Butte, the northern, is a chimneylike plug of

columnar trachybasalt atop a cone of tuff-breccia and lava which had been emplaced in a valley in the former scarp of the Chuska Mountains. On the east flank of the butte Chuska sandstone underlies tuff-breccia at an elevation of almost 8800 feet and forms a ridge between the butte and the mountain front to the east. On the north and west flanks, shelves cut on Chuska sandstone and capped by thin tuff and lava incline about 5° W. toward the lowland north of Wheatfields Creek.

Two miles southeast is an unnamed butte, here called South Tsailee Butte, formed by a tuff-breccia crater filling capped by thick flows of agglomerate and lava and cut by a thick cone sheet of minette on the north flank.

Northeast of Tsailee Butte a thin sheet of lava and tuff caps a headland of the Chuska Mountains.

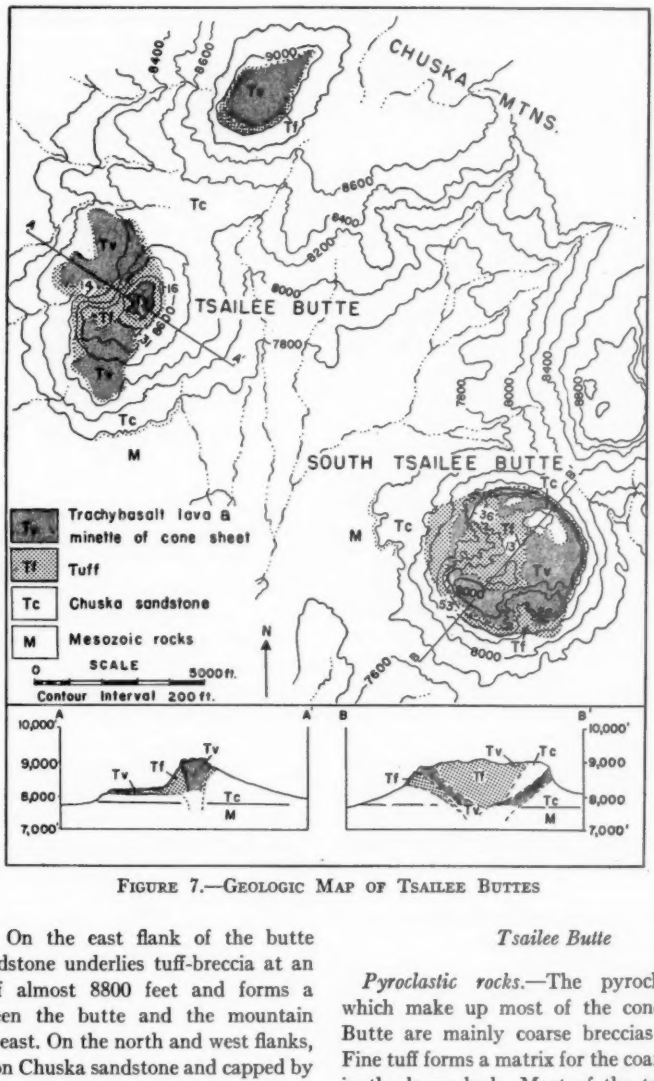

FIGURE 7.—GEOLOGIC MAP OF TSAILEE BUTTES

Tsailee Butte

Pyroclastic rocks.—The pyroclastic rocks which make up most of the cone of Tsailee Butte are mainly coarse breccias and lapilli. Fine tuff forms a matrix for the coarser material in the lower beds. Most of the tuff is sandy.

A ravine cut into the west side of the cone reveals a thick section of pyroclastic deposits and small discontinuous lava flows and agglomerates. The thin discontinuous sheet of tuff that overlies Chuska sandstone on the shelves near the cone is cross-bedded and con-

tains many lenses of sand and pebbles. Small fragments of lava, acidic igneous rock, sandstone, and shale are numerous.

The tuff which caps the shelves is thicker and coarser toward the center of the cone. Near the center, below the plug of lava, coarse igneous breccia and tuff-breccia form a ridge. Broken masses of scoriaceous to dense lava, fragments of tuff and breccia, and xenoliths of sandstone, shale, and acidic igneous rock are cut by apophyses of vesicular to dense trachybasalt. The mass is unbedded, but downslope toward the shelves the breccias become bedded and dip outward from the cone.

Underlying the igneous breccia and coarse tuff-breccia is more than 200 feet of tuff and lapilli in cross-bedded layers 1–10 feet thick. These are interbedded with and contain layers of sand and sandy tuff. Fragments of lava and xenoliths of sandstone, shale, and igneous rock are contained in the tuff-lapilli layers.

On the east flank of the butte the plug intrudes a sheath of tuff-breccia, 100–150 feet thick, dipping inward at 15°–45°. The breccia overlies white to buff reworked sandstone overlying truncated cross-bedded layers of Chuska sandstone.

Lava.—The lava of Tsailee Butte is gray to greenish-gray fine-grained porphyritic trachybasalt. On the shelves north and west of the cone the lava is 10–20 feet thick. It thickens to about 50 feet upslope. Near the top of the cone, discontinuous layers of agglomerate, rubble, and lava overlie the bedded tuff and lapilli-breccia and abut the ridge of coarse tuff-breccia.

The butte is capped by a plug 200 feet high of columnar trachybasalt. The vertical jointing shows slight curvature outward at the base of the butte, similar to the structure of Cabezon in the Mt. Taylor field (Hunt, 1938, p. 68). Though the butte could not be climbed, talus at the base indicates that the dense fine-grained trachybasalt of the columns grades upward to rubbly scoriaceous lava at the top. Johnson (1907, p. 312) reported similar gradation in the Mt. Taylor field. Vertical contacts of the plug with the sheath of tuff-breccia show its intrusive nature.

Sequence of events.—The volcano of Tsailee was opened into a valley in the scarp of the Chuska Mountains. The first ejecta were fairly fine-grained and were reworked by streams and slope wash between eruptions. The thickest sections of cross-bedded sandy deposits revealed by the ravine in the west side of the cone probably mark the line of the old stream valley.

The early stage of volcanism, marked by explosive eruptions separated by periods of quiescence, was followed by increasingly violent eruptions as shown by the progressively coarser debris higher in the cone. Viscous agglomerate and lava were frequently extruded at the end of the volcanism.

At the end of the period of volcanic activity the cone stood somewhat higher than the present top of the mountains. The plug rises to about 9100 feet and probably solidified in the conduit below the summit crater. The thin lava cap underlain by tuff on the headland northeast of the butte indicates that lava flowed eastward from the cone.

South Tsailee Butte

Pyroclastic rocks.—The layers of tuff, lapilli, and coarse breccia that make up most of the butte are generally well sorted, contain thin discontinuous lenses of fluvial sand and gravel, and are locally cross-bedded. Blocks up to 12 inches in diameter of lava, tuff, and red sandstone and shale, and smaller xenoliths of biotite granite and hornblende granite are scattered throughout the sequence.

The pyroclastic rocks form two sequences separated by an angular unconformity. The lower unit, composed of buff to light-blue-gray tuff and lapilli breccia, forms the lower western scarp of the butte and dips inward at angles of 50°–75°. The overlying unit of somewhat coarser and darker lapilli breccia has inward dips of 5°–15°. Both units are sheared and have slumped inward toward the center of the butte. Numerous small discontinuous faults have maximum displacements of 12 inches, and drag folds have been locally developed by sliding in incompetent layers.

Lava.—Overlying the pyroclastic rocks on the southern half of the butte is 300 feet of discontinuous interlayered flows of agglomerate and lava. Large patches of reddish rubble and blocks of lava are churned with scoriaceous trachybasalt in thick masses and tongues of flow breccia. The lava, a dark-gray to greenish-

black fine-grained porphyritic trachybasalt, is exposed in steep cliffs on the west and south sides of the butte.

Cone sheet.—The dike that crops out on the northwest side of the butte is part of a thick cone sheet, a low-angle inward-dipping arcuate dike, cutting the tuffs of the core of the butte. It is about 35 feet thick where it is exposed on the northwest side of the butte, and thickens to a maximum of 250 feet on the east side. It dips inward 20°–40° on the north and east sides. On the west side the dike is more vesicular and rubbly and is apparently continuous with the lava and agglomerate which cap the south half of the butte.

The dike shows layering that is attributable to intrusion. At the lower contact a zone 5–8 feet thick has a gneissoid appearance and weathers to thin plates parallel to the contact with the overlying tuff. The upper contact zone has a sharp basal separation from the coarse blocky-jointed core of the dike, in contrast to the lower zone which has a vague uneven upper contact with the core.

Balk describes a long arcuate dike at Buell Park in the Zilditloi Field as a ring dike, but notes that the dip in one portion is inward, as in a cone sheet, rather than vertical (Allen and Balk, 1954, p. 108).

Crater sandstone.—Two patches of sandstone on top of the butte have a puzzling relationship to the volcanic rocks that completely surround them. They were at first thought to be remnants of a crater filling washed from nearby cliffs of Chuska sandstone, but their lack of contamination by volcanic fragments and their characteristic cross-bedding and cementation indicate that they are pure Chuska sandstone. They are hardened and reddened near the contacts with the intruding cone sheet, and are locally overlain by tuff and lava.

These two patches are believed to be roof pendants of Chuska sandstone into which the cone sheet was intruded. The cone and crater filling of pyroclastic materials protected the sandstone during most of the period of volcanism. Intrusion of the cone sheet into the pyroclastic materials and sandstone placed them in a funnel of resistant lava which has since protected them from erosion.

Sequence of events.—The first stage of volcanism at South Tsailee Butte probably produced a shallow crater in a valley on the flank of the Chuska Mountains. Explosive eruptions built a low pyroclastic cone about the crater. Erosion of the cone and nearby sandstones washed and sorted the pyroclastic materials and contributed sand to the sediments.

At least one interval of subsidence occurred during cone building; probably it was simple inward slumping of part of the crater wall as a result of load and disturbance in the soft, unconsolidated pyroclastic deposits. The fact that the fractures and shear zones are healed, discontinuous, and disappear into unfractured material attests to the unconsolidated nature of the pyroclastic deposits at the time of slumping.

As in the other volcanic areas, quiet extrusion of lava indicated waning of volcanic activity. The volume of lava was small, much less than at Tsailee Butte, because no remnants are preserved except for those capping the butte. Perhaps the crater became plugged, and much energy was expended in the intrusion of the cone sheet.

Roof Butte

General Statement

At the head of Tsailee Creek, on the divide between Lukachukai and Red Rock Valleys, are two volcanic necks and a remnant of a lava flow forming the highest summit of the Chuska Mountains (Figs. 2, 8). The highest and easternmost peak, rising to an elevation of almost 9600 feet, is a circular neck of pyroclastic rocks. The western peak is vent agglomerate and breccia overlain by flow rock. The lava remnant forms a lower mound between the two volcanic necks.

East Peak

East Peak is composed of highly comminuted vent agglomerate. Fragments of Meozoic and Tertiary sandstone and a few Precambrian (?) igneous xenoliths are enclosed in a churned matrix of tuff and lava cut by many apophyses of minette. Surrounding and in part underlying the agglomerate are buff tuff and lapilli-breccia that grade outward and downward into sandy

tuff. Irregular faulted beds of tuff have dips which range from 45° to vertical. Inward slumping of peripheral beds is shown by innumerable small faults and by variable dips.

Sequence of Events

The Roof Butte area shows only minor igneous activity; only a small quantity of ash

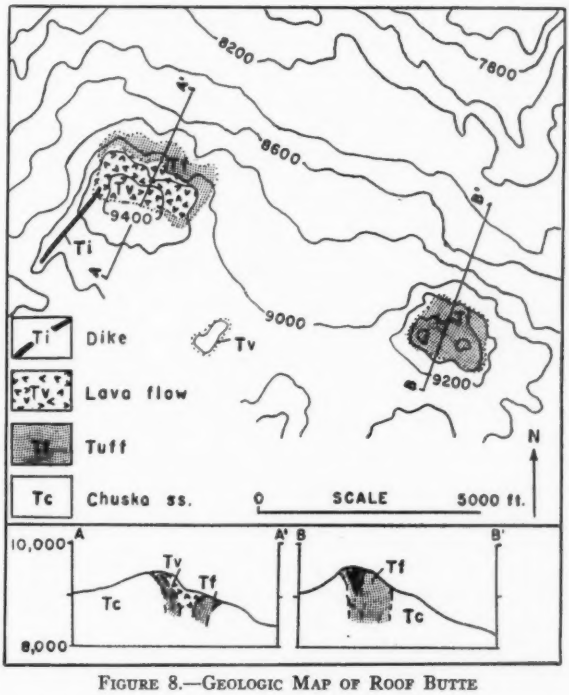

FIGURE 8.—GEOLOGIC MAP OF ROOF BUTTE

West Peak

West Peak is capped by about 100 feet of weathered reddish-brown to grayish-brown fine-grained scoriaceous lava surrounded by dense to vesicular blue-black fine-grained trachybasalt. Erosion of the neck has produced a section cut downward through the core and shows a gradation from flow rock above to vent agglomerate below. The flow rock includes slabs, blocks, and fragments of sandstone, tuff, and earlier flows. The vent agglomerate is a comminuted mass similar to that found at East Peak. The lava is underlain by 5–30 feet of sandy tuff and lapilli-breccia. At the southern edge of the neck calcareous tuffs have dips of 20°–50° inward and show a slumped and somewhat graded contact with the underlying Chuska sandstone. As at East Peak, the tuff and underlying sandstone show evidence of inward slumping.

and flow rock was extruded. The low dips of the bedded rocks at the margins of West Peak and the marked funnel shape of the crater fillings show that erosion has not removed much material since the time of volcanism. Probably initial drilling of the conduits by phreatic explosion produced small craters. Succeeding explosions enlarged the craters, induced inward slumping of the crater walls, and built low cones about the craters. Quiet extrusion of lava ended the volcanic activity.

It is doubtful if the lavas extended much beyond the distance of the one remaining flow remnant south of West Peak, for this is only about 20 feet thick. The dike extending southward from West Peak was probably intruded at the end of volcanism, as it extends upward into the lower portion of the agglomerate in the crater. The rocks of both the flow remnant and the dike are fine-grained green- to blue-black

minette, similar to the trachybasalt of West Peak crater.

General Considerations and Conclusions

Introduction

As a group these five volcanic centers in the Chuska Mountains show similar eruptive history and structural development. They began with violent explosions from vents within the Chuska Mountains of that time. The craters were filled in part with pyroclastic materials deposited directly from the explosions, but to a greater extent by fluvial pyroclastic materials contaminated and interbedded with sands washed from adjacent hill slopes. In most of the vents the crater deposits slumped inward during and after volcanism. After the explosive phase, lava was extruded in low domes and thick flows, and the entire mass was locally intruded by dikes and plugs.

The size of the volcanic remnants reflects the size of the craters or valleys of deposition and the thickness of the contained deposits after collapse, the thickness and erosional resistance of the lava flows that cover the pyroclastic deposits, and the position of the volcanic center in relation to the mountain topography and structure. Since the end of the volcanic episode, dated as late Miocene or Pliocene (Lance, 1955), erosion has stripped 1500–2000 feet of sedimentary rocks from the edges of the Chuska Mountains, and now Sonsela Buttes, Tsailee Buttes, and Shiprock and other necks stand isolated above the erosion surface. Lava remnants at low levels west of Sonsela Buttes and at the Palisades suggest that even the old valley floors were at least 200 feet higher than at present and have thus experienced topographic reversal.

Petrography

General statement.—The volcanic rocks of Chuska Mountains are mainly minette or its extrusive equivalent trachybasalt. (*See* Williams, 1936, p. 148 for comprehensive discussion of petrography of Navajo rocks.) They are mixtures of potassium feldspar, pyroxene, biotite, and accessory ores and apatite. Olivine, usually highly altered, is in many rocks an essential constituent.

Tuff.—The pyroclastic rocks are lithic or between lithic and vitric in Pirsson's classification (1915, p. 193). Rock fragments and, in the finer sizes, single crystals of pyroxene and biotite make up most of the igneous constituents. There is much included quartz, derived from the Mesozoic and Tertiary formations through which the conduits were drilled. Shards of brown and green glass are present in most rocks but nowhere compose more than 30 per cent. The rock fragments include trachybasalt, granite and other acidic igneous rocks, and sandstone and shale.

A Rosiwal analysis of tuff from Washington Pass gave the following average percentages: glass, 30; feldspar (orthoclase, sanidine, microcline, and plagioclase) 30; pyroxene, 10; biotite, 10; quartz, 5; and rock fragments, 15.

Lava.—The trachybasalts of the volcanic centers are predominantly gray to green fine- to coarse-grained rocks in which phenocrysts of pyroxene, biotite, and olivine are visible. The fine-grained lavas, which are generally the later flows, are light colored. The color changes to dark green with increase in grain size.

The feldspar is predominantly sanidine, fresh, glassy, and pseudo-uniaxial (Z:C 0–5°), though in some rocks the feldspar is orthoclase. The feldspar occurs in subhedral to anhedral plates up to 1 cm in diameter. It contains poikilitically all other minerals, including minute blebs of a colorless to brown altered material which Williams (1936, p. 154) and Prisson (*in* Gregory, 1917, p. 107) identified as leucite (?). The pyroxene occurs as euhedral, distinctly zoned crystals: the interior is pale-green diopside (?), the rim dark-green aegirine-augite. In some cases pyroxene is present in two generations, and earlier large broken crystals a maximum of 2 mm long and later pale-green unzoned crystals are included in the feldspar. Biotite is also present in two generations. The older biotites are zoned with pale-tan interiors and corroded dark-red rims, whereas the younger crystals are tiny, pale yellow, and highly pleochroic. Olivine is present in small rounded grains as much as 1 mm in diameter and is commonly altered to serpentine (iddingsite?) and iron oxides.

Rosiwal analyses of lavas from Washington Pass and the Palisades gave the following percentages:

	Washington Pass			Palisades
	Flow A Fine-grained	Flow B Coarse-grained	Flow C Fine-grained	Fine-grained Trachy-basalt
Sanidine	25	32		26
Orthoclase			20	
Pyroxene	35	32	14	33
Biotite	6	12	20	
Olivine	9	12	12	14
Ores	6	6	10	6
Apatite	1	2	3	2
Glass and Leucite (?)	18	4	20	19
Rock fragments			1	
	100	100	100	100

Intrusions.—The dike rocks of the Chuska Mountains are dense dark-greenish-black minettes in which phenocrysts of pyroxene and biotite are visible. Olivine is rare. The rocks are mineralogically similar to the trachybasalts, which were obviously fed by the intrusions. They generally form coarse blocky outcrops whose platy borders are parallel to the contacts.

Metamorphism

Metamorphic effects on country rock and xenoliths are confined to local reddening or silicification, probably because of lack of mineralizers. At Tsailee Butte are xenoliths of Mississippian (?) limestone which were apparently unaffected by heat and retain identifiable brachiopod and pelecyod fossils in a dense clastic limestone matrix.

The tuff underlying the lava is cemented with calcite or, especially at Washington Pass, with silica, both probably derived from the volcanic rocks. At the Palisades, shear zones and fractures are filled with veinlets of calcite, and thin sections of the underlying Chuska sandstone show pre-existing opal and chalcedony cement replaced by calcite.

At Black Pinnacle Williams (1936, p. 158) found the intruded sandstone converted to buchite but concluded that vitrification could take place at low temperatures. At Fluted Rock Balk noted some buchite formed in xenoliths of sandstone and siltstone, but also found fossil wood unaffected by heat (Allen and Balk, 1954, p. 106).

Bedded Crater Deposits

The explosion pipes of the Chuska Mountains are characterized by fluvial infillings of interbedded sand and pyroclastic materials. Descriptions of such well-bedded crater deposits are relatively uncommon in geologic literature. Wagner (1922, p. 27) describes the Pretoria salt pan in South Africa as an explosion pipe filled with interbedded pyroclastic deposits and lacustrine clays. In the Hopi Buttes area, Hack (1942, p. 350) describes similar fillings in craters of diatremes. Daly (1925, p. 26) attributes bedded deposits in a vent on Ascension Island to fluvial infilling from outside the crater.

In the Chuska Mountains the explosion pipes are filled with well-stratified sand and tuff, in contrast to the lacustrine infillings of Pretoria and the Hopi country. In those areas the craters were built in aggrading basins dotted with lakes and swamps, where crater lakes and streams dammed with pyroclastic material allowed deposition of clay, silt, and evaporite from the quiet calcium-charged waters. In the Chuska Mountains the craters were opened in valleys on the crest and flanks of the mountains. They were probably quickly filled with sand and tuff eroded from the bordering steep slopes, and lakes had no opportunity to form. Only in Washington Pass, where a thin limestone is locally interbedded with the fluvial deposits, would there have been a crater lake.

In further contrast, Balk (Allen and Balk, 1954, p. 107–122) makes no mention of fluvial deposition for the much more faintly bedded lapilli tuff that constitutes the Green Knobs and parts of Buell Park, respectively only 6 and 10 miles west of the Chuska Mountains in the Zilditloi field. Nor is there any interbedding of fluvial sand here or any contamination of tuff by sand or other particles that cannot be considered xenoliths from the volcanoes themselves. Perhaps these centers were too far from the rough topography of the Chuska Mountains of the time, so that there were no adjacent sandstone slopes to furnish stream sediments to the craters.

Deformation of Crater Deposits

Deformation of the crater deposits is indicated by fractures, shear zones, and locally drag folds. Most of the deformation was probably caused by slumping and sliding of soft unconsolidated sediments because of load and oversteepening of crater walls by explosions. This mechanism is illustrated by the crumpling and shearing in the tuff at the base of the lava remnant at the southeast corner of the Palisades. The deposits have slid downward, presumably because of the steep slope (40°–60°) and the load of the overlying lava.

Angular unconformities in the deposits at South Tsailee Butte and Washington Pass suggest that explosions removed parts of the crater walls during periods of predominantly fluvial deposition. The unconformities are local and disappear within a quarter of a mile. The surfaces of the unconformities are not channeled as if streams had breached the crater. The breaks are overlain by coarse explosion breccias interbedded with fluvial deposits. Explosion breccias interrupting sequences of fluvial deposition are common in the craters and show that vents were active even during deposition. Blocks of lava that dropped into fluvial deposits deformed the underlying sedimentary rocks. In the Palisades such deformation affected the sedimentary rocks as much as 10 feet away, as indicated by fractures radiating outward from the block.

There is no conclusive evidence of large crater subsidence in most of the vents of the Chuska Mountains as Hack (1942, p. 365) has inferred in the Hopi Buttes diatremes, where faulting, steeply-dipping beds, and basin-shaped crater deposits suggest subsidence into the conduit. South Tsailee Butte, Roof Butte, and the vent near East Sonsela Butte have steeply-dipping beds and evidence of small-scale slumping in the crater deposits, but this may have been caused by inward surficial sliding after explosions. The uppermost beds of tuff at South Tsailee dip inward only 13°, not steeply enough to indicate deep central subsidence. The small amount of subsidence indicated at most of the centers is comparable to that recorded at the Green Knobs in the nearby Zilditloi field by Balk (Allen and Balk, 1954, p. 111), who, however, in addition infers from the joint pattern some upthrusting of the plug. At Buell Park, on the other hand, Balk concluded that the tuff plug had moved neither up nor down.

Washington Pass, in contrast to the other centers, shows evidence of caldera subsidence—steeply dipping rim rocks, folding in the lava and tuff, faults in the underlying bedrock. These features require removal of magma at depth, probably from an enlarged conduit rather than collapse of a block overlying the magma chamber as is postulated in cauldron subsidence accompanied by ring faults.

Magma Reservoir

The distinctive petrography of the entire Navajo-Hopi volcanic province indicates that the centers were supplied by a common magma reservoir. Williams (1936, p. 171) suggests that the reservoir was shallow on the basis of the highly explosive character of the volcanoes and the close spacing of the vents. Xenoliths of Precambrian rocks in most of the pyroclastic deposits demand that the reservoir be at least as deep as the basement, and Williams suggests that the volcanoes were fed from a silllike body at the contact of Archean and overlying rocks. Although the basement rocks are exposed near Fort Defiance and are perhaps 1500 feet deep at the Palisades, they are at a depth of 9000 feet near Shiprock (McKee, 1951, pl. 3).

The volcanic centers of the Chuska Mountains area itself are aligned approximately along the Defiance monocline. This early Tertiary structure is probably an upward reflection of a major fracture formed by compression in the underlying basement (Baker, 1935, p. 1500; Kelley, 1955, p. 799). The volcanism may be related to later tensional stresses along this fracture.

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MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY MARCH 9, 1956

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 68, PP. 469-480, 7 FIGS. APRIL 1957

PALEOCURRENTS OF LAKE SUPERIOR PRECAMBRIAN QUARTZITES

By F. J. PETTIJOHN

ABSTRACT

Data on the attitude of the cross-bedding in the Sioux, Baraboo, Waterloo, Barron, Sturgeon, Mesnard, Mississagi, and Lorrain quartzites, all of Precambrian age, are summarized. The azimuth and inclination of the cross-bedding, corrected for tilt of the beds, show a regional uniformity and a remarkable persistence through time. The prevailing southeasterly dip of the cross-laminations shows that these deposits were derived from a highland north and northwest of their present outcrops.

The inclination of the cross-laminations, tilt-corrected, commonly exceeds the angle of repose. Inasmuch as the corrected mean inclination in a given area is closely correlated with the mean dip of the strata and varies also with the position of the cross-strata on a fold, the cross-laminations are vectors rotated by internal shear of the quartzite during folding.

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INTRODUCTION AND ACKNOWLEDGMENTS

Mapping of cross-bedding and other primary current structures has proved useful in reconstruction of regional paleoslopes (Potter and Olson, 1954, p. 72), fixing the trend of the ancient shore lines (Tanner, 1955), and locating probable source areas of the sands forming the cross-bedded strata (Potter and Siever, 1956). It seems appropriate, therefore, to examine and map primary vector properties of the Precambrian quartzites of the Lake Superior region. Preliminary work by the author and his students (Trow, 1948, Ph.D. thesis, Univ. Chicago; Brett, 1955) was encouraging, so that in 1954 more extensive work was undertaken.

The principal achievement was the reconnaissance mapping of cross-bedding and associated structures in the Lorrain quartzite of Huronian age in the north-shore district of Lake Huron—

the type Huronian area. A few observations were also made on the Mississagi quartzite of the same region and on the Mesnard quartzite near Marquette, Michigan. These data were combined with earlier observations, and a generalized paleocurrent map of the Precambrian quartzites of the Lake Superior region was compiled (Fig. 1).

The writer was assisted in the field by John Schlee and P. W. Choquette. A grant from the Penrose Bequest of The Geological Society of America is gratefully acknowledged. The writer is also indebted to Paul E. Potter for many discussions of the project both in the field and in the office, and to George W. Brett and Raymond Siever for unpublished data on the Sioux and Barron quartzites, respectively.

Lorrain Quartzite and Related Strata

The Lorrain is a widespread formation, primarily a medium- to coarse-grained quartzite of Huronian age (Collins, 1925, p. 67). It belongs to the Cobalt Series; the type locality is the township of Lorrain, Cobalt district. The Lorrain occurs in several isolated areas (Collins, 1935). Westernmost is the Bruce Mines area where the Lorrain is well exposed and conglomeratic. A second area, in which more sharply folded Lorrain is present, is the Lacloche Mountains district on the north shore of Lake Huron between Espanola and Little Current, Ontario. The third and most extensive area, in which the formation is the least deformed, is that found at Cobalt, Ontario, and environs. The writer restricted his observations to the Bruce Mines area and to the vicinity of Whitefish Falls in the Lacloche Mountains.

The Lorrain is one of the thickest quartzite formations known to the author. It is nearly 6500 feet thick in the Lacloche Mountains and is 5500–6000 feet thick in the Bruce Mines district (Collins, 1925, p. 69). In the latter area the lower half of the formation is pink to reddish; the top half is white. Scattered through the section are pebbly or conglomeratic beds characterized by well-rounded pebbles of white vein quartz and brilliant red jasper. The jasper-bearing conglomerates are well known to North American geologists, as specimens from the outcrops or the glacial deposits derived therefrom are in almost all college and university rock collections. Only the lowest Lorrain is appreciably feldspathic; most of it is highly quartzose, commonly strongly cross-bedded, and uniformly quartzite or conglomeratic quartzite. No slate or other rocks were found within the Lorrain.

Cross-bedded (current-bedded or false-bedded) strata are common, and only a few outcrops fail to show one or more such layers. The cross-bedding is what is commonly called torrential or continuous incline-bedded banks (Anderson, 1934); this type has also been described as diagonal false-bedding (Richey and Kennedy, 1939) and as planar cross-bedding (McKee and Wier, 1953, p. 387).

Methods of Investigation

As the cross-laminated sets were generally bounded by plane lower and upper surfaces, one could measure and record (1) strike and dip of the cross-bedded unit, (2) strike and dip of the cross-laminations, and (3) thickness of the cross-bedded layer. Most of the cross-laminations were plane surfaces; few were curved surfaces, tangential to the base of the bed, so that there were no difficulties in ascertaining both strike and dip. It was necessary to find an exposed surface of a cross-lamination or to see traces of the laminations on two intersecting surfaces, in order to measure the true position of the lamination planes in space. The thickness of the cross-bedded layer was measured at the place where the strike and dip of the contained laminations were recorded.

Two measurements of strike and dip of the cross-laminations per bed were made; measurements were made on two to five beds per outcrop. Outcrops were no less than a mile apart, and in general two to five outcrops were studied per 6-mile interval ("range") along the outcrop belt in the Bruce Mines area. Outcrops are numerous along the available access roads; 119 cross-bedding measurements were made on the Lorrain in the Bruce area, and about 32 were made in the Whitefish Falls area.

Inasmuch as the strata studied have been folded and are locally vertical, or even overturned, it was necessary to correct the readings on the cross-laminations for the tilt of the beds. This was done by rotating the pole of the cross-bed, plotted on a stereonet, through the

FIGURE 1.—PALEOCURRENT MAP OF PRECAMBRIAN QUARTZITES OF THE LAKE SUPERIOR REGION

Areas shown in black are principally quartzite. Mean-current azimuth shown by arrows. Figures denote number of measurements on which mean is based.

angle of dip of the bed itself about the strike of the bed. The writer assumed that the azimuth and inclination of the cross-bedding thus corrected are essentially the same as at the time of deposition. That this assumption is valid, despite some theoretical objections (Whitaker, 1955, p. 765), has been demonstrated by Kopstein (1954, p. 64) in the Harlech dome area of Wales. The general consistency of the data for the Precambrian strata studied seems also to justify this procedure.

The readings, corrected for tilt, have been summarized in circular histograms (Fig. 5), and the mean cross-bedding azimuth has been plotted (Fig. 1). The summarization of circular data presents some difficulties which have been reviewed by Tanner (1955) and others (Griffiths, 1953; Jizba, 1953; Chayes, 1954). The measurements made by the author and associates, however, yield unambiguous distributions, so that the pattern revealed by graphic plot (histogram or map) can readily be discerned with the naked eye and requires no statistical analysis to demonstrate its existence or validity. The arithmetic mean and standard deviation of the cross-bedding azimuths, inclinations, and thickness were computed and tabulated. No geologically significant difference in mean azimuth would have been obtained, no matter what method of summarization was used.

Cross-Bedding

Thickness (Scale)

Twenty-seven measurements were made of the thickness or scale of the cross-bedded units. The thickness of the sedimentation units ranged from 4 to 55 inches. The average thickness was 16.7 inches. Schwarzacher noted (1953, p. 324) that the thicknesses of cross-bedded units are log normally distributed. As the data on the Lorrain quartzite are meager no attempt was made to test the thickness distribution for normalcy.

The significance of the scale of the cross-bedding is not known, although several authors have discussed this property (Schwarzacher, 1953, p. 325). Scale may be correlated, in part, with coarseness of grain; silts exhibit "small-scale" cross-bedding, whereas gravels show "large-scale" torrential bedding. But even sands show a wide range in scale of cross-bedding, and the factors that control this feature are not known. The cross-bedding of the Lorrain is similar in scale to that observed in the other Lake Superior Precambrian quartzites (Table 1). More field observations on thicknesses of cross-bedded units should be made, and some understanding of the origin of cross-bedding is needed before this attribute can be properly interpreted.

Table 1.—Scale of Cross-Bedding
(In inches)

Formation	Mean	Maximum	Minimum	No. measurements
Lorrain	16.7	55	4	27
Baraboo*	less than 12	36–40	1	...
Sturgeon	8	17	3.5	10
Mesnard	7	18	2.5	15
Moine†	6±	rarely exceed 12	...	...
Torridonian†	may exceed 24			...
Lower Greensand**	10	100	2.5	100

* After Brett (1955)

† After Wilson, Watson, and Sutton (1953, p. 378)

** Estimated from graph of Schwarzacher (1953, Fig. 3)

Inclination

For measurement and analysis the cross-laminations are considered planes. Although in theory cross-laminations become tangent to the base of the bed and are truncated by the top surface, in fact such concavely curved surfaces are rare; most cross-laminations are substantially plane surfaces. The inclination of the cross-beds can be defined, therefore, as the dihedral angle between the plane of the cross-lamination and the plane of the true bedding.[1]

[1] If the initial dip of the bed is not zero, the inclination thus defined is not the angle of repose. As the difference in these angles is small and as the strata are tilted and the original dip unknown, the inclination is treated as though it were the angle of repose.

In the Precambrian quartzites studied, the inclination varies widely. In the Lorrain quartzite of the Bruce Mines area it ranges from 6° to 41° and averages 20.2°. In the Lacloche area the cross-laminations in the Lorrain were inclined 5°–59° and averaged 25°. Studies of other Precambrian quartzites yield similar values (Fig. 2).

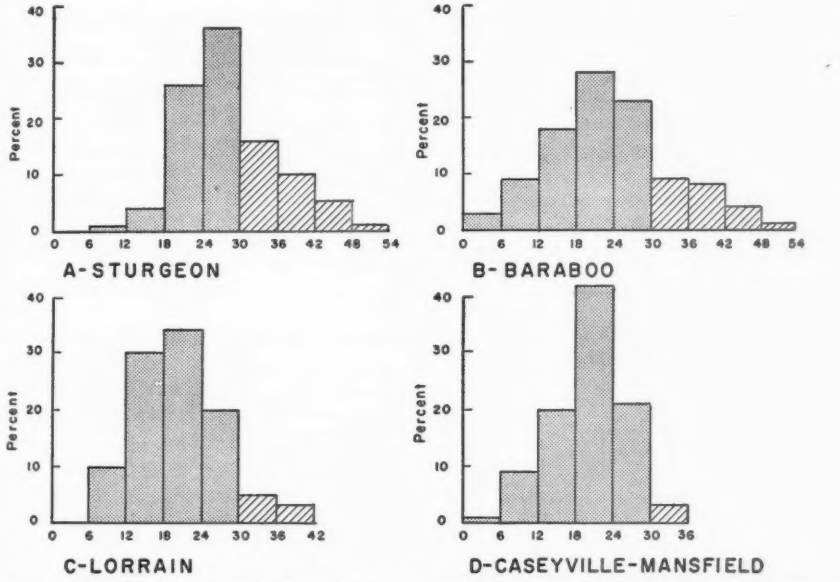

FIGURE 2.—HISTOGRAMS SHOWING CROSS-BEDDING INCLINATIONS IN LAKE SUPERIOR PRECAMBRIAN QUARTZITES AND OTHER FORMATIONS

A, Sturgeon quartzite, Michigan (after Trow, 1948, Ph.D. thesis, Univ. Chicago); *B*, Baraboo quartzite, Baraboo, Wisconsin (after Brett, 1955); *C*, Lorrain quartzite, Bruce Mines, Ontario; *D*. Caseyville-Mansfield (Pennsylvanian), Illinois-Indiana (after Potter and Olson, 1954). Classes near or above angle of repose shown by diagonal ruling.

The inclination must be a function of the angle of repose of the sand at the time of deposition. The angle of repose of dry sand has an upper limit of 34° (Bagnold, 1942, p. 210). The angle is said to be higher for aeolian sands than for subaqueous deposits. McKee (1940) for example, measured cross-bedding in formations in the Grand Canyon region and found maximum angles of 27°–28° in the two water-laid sandstones and 33° in a sandstone considered aeolian. These observations seem to be supported by experiments which show that the same sand has a lower angle of repose in water than it does in air (McKee, 1955).

Examination of measurements of Precambrian cross-laminations, however, show many anomalously high inclinations. Many greatly exceed the angle of repose, even of dry sand. The most reasonable interpretation is that the laminations have been rotated by deformation. Although there is some evidence of soft-sediment movement and slump, most of the rotation must be due to tilting of the strata, inasmuch as few measurements of cross-bedding in untilted formations show inclinations exceeding the angle of repose (Fig. 2, *D*).

Deformation might decrease as well as increase inclination of cross-beds, so that the mean inclination would remain unchanged. This does not seem to be the case, as the mean inclination of the cross-laminations of the quartzites is consistently greater than that of undeformed strata (Table 2), and the mean inclination appears to show a regular increase

with increase in the mean dip of the quartzite beds.

It is not clear how deformation by folding produces the higher inclinations, and why the considerations. Brett (1955) made a similar analysis of the cross-laminations in the Baraboo quartzite and noted that the mean inclination of the cross-beds in the north limb of the

TABLE 2.—RELATION OF MEAN DIP OF BEDS TO MEAN INCLINATION OF CROSS-BEDDING

Formation and locality	No. measurements	Mean dip	Mean inclination	Reference
Pennsylvanian (Illinois and vicinity)	531	0°	16.5°	Potter and Olson (1954, p. 58)
Lorrain quartzite (Bruce Mines)	119	19°	20.2°	This paper
Baraboo quartzite (Wisconsin)	283	46°	23°	Brett (1955)
Lorrain quartzite (Lacloche)	32	77°	25.8°	This paper
Sturgeon quartzite (Michigan)	80	ca 90°	28.1°	Trow (1948, Ph.D. thesis, Univ. Chicago)

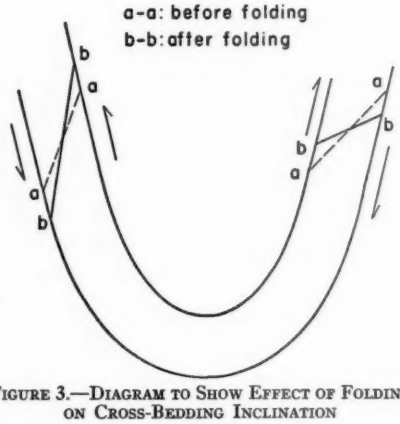

FIGURE 3.—DIAGRAM TO SHOW EFFECT OF FOLDING ON CROSS-BEDDING INCLINATION

mean inclination increases. If the fold is a flexure fold one can postulate internal shear of the "card-deck" variety parallel to the plane of the quartzite bed. If the cross-laminations are suitably oriented such shear will rotate the laminations and produce a change in inclination. The effect should be dependent on the position of the cross-laminations on the fold, and the rotation should increase the inclination on one limb and decrease it on the other (Fig. 3). To check this concept, the inclinations of the cross-laminations on the two limbs of the Lacloche syncline were separately averaged. The mean inclination of the cross-beds on the north limb was 17.9°, whereas that on the south limb was 32.0° (Table 3). The difference was as would be predicted from the above Baraboo syncline was 20°, whereas that of the south limb was 24.5°. Again the data are consistent with theory, but the difference in inclination on the two fold limbs is less than that in the Lacloche area, perhaps because the Baraboo structure is a more open fold. The inclinations of the cross-laminations in the northward- and southward-dipping beds in the Bruce Mines area do not appear to be significantly different. As the mean dip of the beds in this area is much less than in either the Baraboo or Lacloche areas, however, the effect of the deformation on the cross-lamination might be very small.

Elementary analysis of the geometry of the card-deck shear parallel to the bedding of the quartzite will show why the mean inclination appears to increase with increasing deformation if the inclination is reduced on one fold limb and increased on the other (Fig. 4). A unit of shear will rotate the inclination through a lesser angle, if the cross-lamination is inclined in one direction with respect to the shear couple, than if it is oriented in the opposed position.

It is not certain, however, that folding is of the flexure type and that the shear is on planes parallel to the true bedding. If the fold is a shear fold and characterized by a flow cleavage, the effect on cross-bedding inclinations is less simple and not so easy to predict. Presumably the inclination should be steepened where the beds are thickened and diminished where the beds are attenuated. Only in the apical areas

would the cross-beds be abnormally steep. As fold limbs are more common than apical areas the observed average inclination of cross-beds should be diminished. As these relations were not observed most of the folds examined are flexure folds.

TABLE 3.—RELATION BETWEEN POSITION ON FOLD AND INCLINATION OF CROSS-BEDDING

	Lorrain quartzite (Lacloche Mountains)		Baraboo quartzite	
North limb	14 beds	17.9°	115 beds	ca 20°
South limb	18 beds	32.0°	157 beds	ca 24.5°
Mean	32 beds	25.8°	272 beds	22.5°

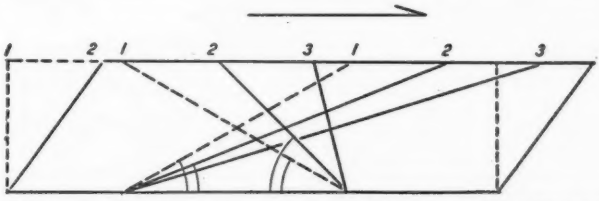

FIGURE 4.—DIAGRAM TO SHOW EFFECT OF "CARD-DECK" SHEAR ON CROSS-BEDDING INCLINATION

However, the mechanics of folding is much more complex than commonly supposed. (*See* Cloos, 1947.) The nature of the movements and the volume relations involved during deformation are very poorly understood. The study of oölites or pebbles contributes much to the analysis of the deformation; but cross-bedding, *Scolithus* tubes,[2] and columnar stromatolites provide a vector, the rotation of which during folding should also contribute to understanding the mechanics of folding. Clearly, too, the maximum inclination of cross-laminations in tilted beds cannot be used to discriminate between aeolian and aqueous deposits.

Azimuth

The principal objective of this work, however, was not study of the inclination of the cross-beds nor of the effect of the deformation of the inclination of this structure; the objective was study of the variations of the dip direction (azimuth) of the cross-beds and their paleogeorgaphic significance.

Figure 5 and Table 4 show that for each quartzite formation studied in each district the cross-bedding has a strikingly preferred orientation (after correction for tilt). In most cases most of the dip azimuths fall in one quadrant. In all Precambrian quartzites of the Lake Superior region the mean direction of current flow was from the northwest to the southeast. Although most of the cross-bedding azimuths lie in one quadrant there is some scatter about the mean direction. The standard deviation is variable (Table 4). That of the Lorrain in the Bruce Mines area is 58°; that of the Sturgeon quartzite in Michigan is 38.5°. That of the Coconino sandstone in Arizona, presumed to be aeolian, ranges from 38° to 64° (in localities with 50 or more observations). In most cases the azimuthal distribution is unimodal and symmetrical about the mean. The bimodal distribution in the Whitefish Falls area—an exception—seems to be due to a sampling bias.[3]

The most striking feature of this study is the regional uniformity of orientation of the current azimuths (Fig. 1). The mean direction in many districts is similar, despite the probability that the quartzites are not all the same age. The depositing currents flowed from northwest to southeast. The mean cross-bedding azimuths of the Sioux, Baraboo, Sturgeon, and Lorrain quartzites are almost identical;

[2] Dale (1902, p. 12) observed *Scolithus* tubes inclined 35° from the vertical in tilted Cambrian quartzite near Arlington, Vermont. As these tubes when formed were vertical they have been rotated through a 35° angle with respect to their original position.

[3] The horizontal rock surfaces examined (Fig. 6) do not show the prevailing cross-bedding (modal class), owing to the orientation of the surface of the outcrop with respect to the cross-bedding. Cross-bedding in the proximal classes is readily seen; hence the units usually seen to be crossbedded and therefore selected for measurement exhibit the proximal rather than the modal directions.

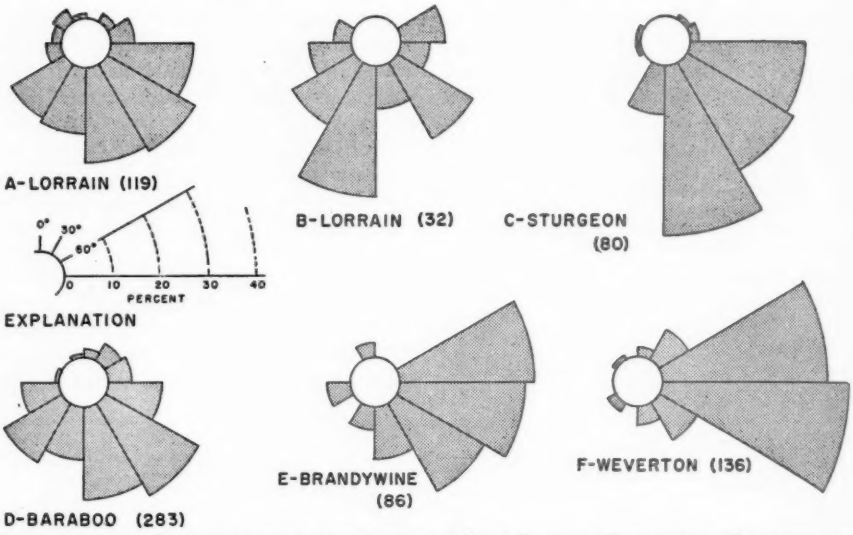

FIGURE 5.—CROSS-BEDDING AZIMUTH DISTRIBUTION IN LAKE SUPERIOR PRECAMBRIAN QUARTZITES AND OTHER FORMATIONS

A, Lorrain, Bruce Mines area, Ontario; *B*, Lorrain, Whitefish Falls area, Ontario; *C*, Sturgeon quartzite, Menominee district, Michigan (after Trow, 1948, Ph.D. thesis, Univ. Chicago); *D*, Baraboo quartzite, Baraboo district, Wisconsin (after Brett, 1955); *E*, Brandywine sand (Pliocene?), Maryland (after Schlee, 1956, Ph.D. thesis, The Johns Hopkins Univ.); *F*, Weverton quartzite (Cambrian) Maryland (after Whitaker, 1955)

TABLE 4.—STATISTICAL SUMMARY OF CROSS-BEDDING AZIMUTHS

Formation and place	Mean	Standard deviation	No. readings	References
Mississagi, Bruce area	100°	.....	15	This paper
Lorrain, Bruce area	163°	58°	119	This paper
Lorrain, Lacloche	177°	.....	32	This paper
Mesnard, Michigan	97.5°	.....	16	This paper
Sturgeon, Michigan	143°	38°	80	Trow (1948, Ph.D. thesis, Univ. Chicago)
Baraboo, Wisconsin	163°	.....	283	Brett (1955)
Sioux, South Dakota, Minnesota, Iowa	162°	.....	39	Brett (unpublished)
Coconino, Arizona	.....	38° to 64°	Over 50 per locality	Reiche (1938)
Streams Georgia coastal plain	103°	42°	64	Tanner (1955)

those of the Mesnard and Mississagi are similar to each other but are more westerly, although the depositing current seems also to have come from the northwest. Data are very meager on these beds and on the Barron quartzite—the only quartzite with a northeasterly rather than northwesterly mean.[4]

[4] Based on four measurements made by Raymond Siever.

The general uniformity in paleocurrent flow over a distance of about 1000 miles is one of the most striking results of this study.[5]

[5] The current azimuth for the Waterloo quartzite is a qualitative estimation by P. E. Potter; that for the Lorrain near Cobalt (Rabbit Lake) is also based on a qualitative statement by Johnston (1954, p. 1055) who observed that "... well-developed cross-bedding indicates a source of sediment to the north."

A second and equally astonishing finding is the stability of the paleocurrent system in time. The measurements for the lower part of the Lorrain in the Bruce Mines area were compared with those of the upper or white Lorrain. The azimuth means were substantially the same.

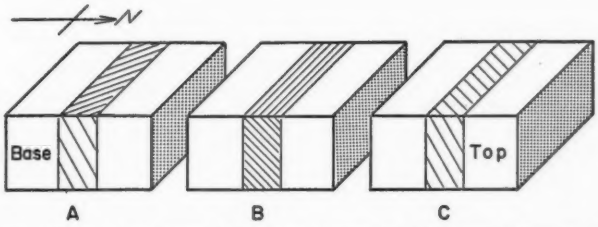

FIGURE 6.—BLOCK DIAGRAM TO SHOW CAUSE OF SAMPLING BIAS AND BIMODAL DISTRIBUTION OF CROSS-BEDDING AZIMUTHS

Thus the current system was maintained without change during the time required to deposit almost 7000 feet of quartzite. A similar uniformity in time has been found to be true of the 5000-foot Baraboo quartzite in Wisconsin (Brett, 1955, p. 146). As the Mississagi is older than the Lorrain and is separated from it by several thousand feet of strata and a major unconformity, it is of interest to note that there appears to be a difference in mean cross-bedding azimuth between this formation and the Lorrain. The shift in azimuth is some 60°–70°, although the current mean is still in the same quadrant. The mean of two measurements on the "upper white quartzite," an unnamed formation about 1000 feet above the Lorrain, is identical with the mean of the Lorrain itself. Thus through almost 20,000 feet of strata the current flow was from a single (northwest) quadrant (Fig. 7).[6]

Comparison of the azimuthal data collected from the Precambrian quartzites with the measurements on cross-bedding in younger formations, especially those of known origin, shows that in general, the scatter or spread of the azimuths is similar to that of the younger beds (Fig. 5, *E*, *F*). As one of these (Fig. 5, *E*) is of fluvial origin, a fluvial origin for the Lorrain is possible, but as cross-bedding data on unquestioned marine deposits are exceedingly meager no conclusions on differences in scale, inclination, or azimuthal variations between aeolian, marine, or fluvial deposits can be made.

The currents responsible for the deposition of the various Precambrian quartzites appear to have flowed from northwest to southeast down a regional slope, from an upland or highland to a lower basin or trough. The uniform regional orientation precludes the possibility of azimuth orientation by littoral currents parallel to a shore line rather than by currents which moved downslope. This downslope flow continued through a long period of time, although a shift in the mean direction of flow occurred between the time of deposition of the Mississagi and the Lorrain quartzites in the Bruce Mines area.

Such regionally uniform orientation and persistence in time strongly suggest a northwesterly source for the sands of the various quartzites. Whatever their age, the quartzites were derived from older rocks which lay northwest of their present outcrops. As the quartzites are probably not all the same age, they may have been derived from different source rocks. The flood of red jasper in the Lorrain and its scarcity in the Mississagi suggest different sources for these quartzites, as perhaps does the shift in mean cross-bedding azimuth. The presence of the jasper in the Lorrain of the Bruce Mines area and its absence in the Lorrain of the Lacloche Mountains and in the Cobalt-Gowganda region suggest a westward or northwestward origin of these materials—a conclusion consistent with the paleocurrent data. These observations are not inconsistent with a derivation of the Lorrain jasper from the jaspilites of the Precambrian of

[6] The uniformity of orientation of cross-bedding is not unique. It has been observed throughout a 12,000-foot sequence in the Moine series and has been characterized by Sir Edward Bailey as "... the most surprising single phenomenon" respecting these strata (Wilson, Watson, and Sutton, 1953).

the Marquette district. If this surmise is correct the Michigan strata must be appreciably older than the Cobalt series.

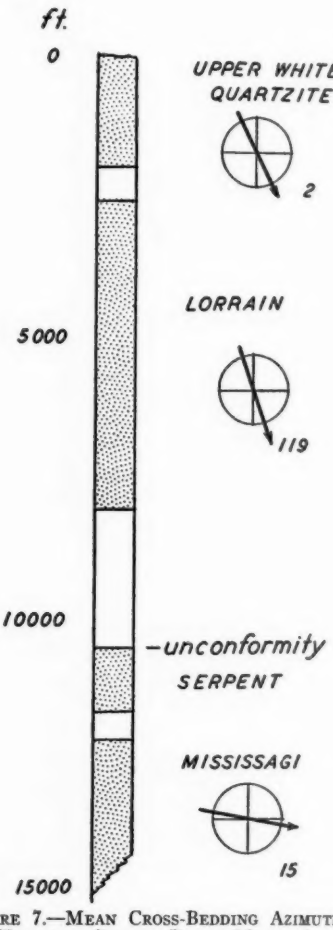

FIGURE 7.—MEAN CROSS-BEDDING AZIMUTHS IN HURONIAN STRATA, BRUCE MINES AREA

Some writers (Cooke, 1933, p. 439) have assumed that the Grenville is pre-Huronian, perhaps formed a crystalline upland during the deposition of the Huronian, and hence was a probable source of the Lorrain and other Huronian sediments (Johnston, 1954, p. 1055). The paleocurrent data preclude the Grenville rocks and Grenville province as the source of the Huronian sediments. The Grenville has also been considered a marine, offshore phase of the Huronian (Collins, 1925, p. 62; Quirke and Collins, 1930, p. 3), and the Huronian has been thought to be derived from a landmass to the north. The paleocurrent data are in harmony with this view; although they indicate a northwesterly source area, they do not prove (or disprove) the equivalence of the Grenville and the Huronian.

Origin

The conditions of deposition of the cross-bedded quartzites and the origin of the cross-bedding remains obscure. The quartzites have been considered both continental and marine. The Mississagi, for example, has been variously attributed to the action of both rivers and beaches (Joubin and James, 1956). The Lorrain is ascribed to a "... shallow submergence" (Collins, 1925, p. 74). The Sioux quartzite has been assigned to "... a marine or lake environment rather than a land environment" (Baldwin, 1949, p. 10). The transition beds between the Gowganda and the Lorrain in the Whitefish Falls area clearly record a progressive shoaling of the waters and a probable transition from subaqueous (probably lacustrine) to subaerial (fluvial) environment.

Knowledge of these formations is incomplete: the quartzites therefore cannot be used to determine the environment of formation of the cross-bedding, nor can the cross-bedding be used to determine the environment of deposition of the quartzites. It is noteworthy, however, that the type of cross-bedding—scale, inclination, and azimuth variations—is similar in all the quartzites, whatever their age or origin.

Even more obscure is the origin of the cross-bedding. Although known to be produced by the wind, as well as by both fluvial and littoral currents, the actual mechanism of origin of the water-formed cross-bedding is not clear. Any mechanism proposed must explain the scale, inclination, and orientation. Some cross-bedding (festoon) appears to fill troughs oriented longitudinally with respect to current flow. The origin of these structures and their fill is obscure, since their formation seems never to have been observed (Stokes, 1953, p. 27). Although festoon cross-bedding was seen in the Lorrain, it is generally very rare. Some authors ascribe the cross-beds to foreset depo-

sition on advancing microdeltas but do not say what these microdeltas are, how or where they originate, or what determines their scale and orientation. Others attribute torrential cross-bedding to formation and migration of barchan-like deposits (Potter and Siever, 1956, p. 231). Some cross-bedding is a product of rippling and is a record of such ripple migration (McKee, 1939, p. 71). Most of the cross-bedding attributed to ripple marking is very small scale. Perhaps the torrential cross-bedding—the most characteristic cross-bedding of the Precambrian quartzites and other sandstones—is the product of the formation and migration of large ripples or megaripples (Kindle, 1917, p. 55). Possibly all cross-bedding is related to the formation and migration of sand waves—large aeolian waves (dunes or aeolian cross-bedding), megaripples or sand waves (common cross-bedding), and ordinary sand ripples (small-scale cross-bedding).

Other Primary Current Structures

Some data were collected on other current structures in the Precambrian quartzites, especially ripple marks. About 22 unpublished measurements by Brett on the ripple marks in the Sioux quartzite showed orientation normal to the mean-current azimuth determined by study of the cross-beds. The same was true of the 21 ripple marks measured in the Baraboo quartzite (Brett, 1955, p. 147). Measurements on about 200 ripple marks in the Sturgeon quartzite, however, showed that only a few were so oriented; the majority had a strong preferred orientation more or less parallel to the current systems deduced by study of the cross-bedding (Trow, 1948, Ph.D. thesis, Univ. of Chicago). Most of the ripple marks in the Sturgeon, however, were symmetrical or were oscillation ripples unlike those in the Baraboo which are asymmetrical current ripples. Ripple mark is very rare in the Lorrain. It was found only near the base of this formation in the transition beds between the Gowganda and the Lorrain. As the Gowganda is a thinly varved argillite, the transition beds were subaqueous, although of shallow-water origin. The Lorrain appears to be of subaerial origin, probably fluvial, and hence has few preservable ripple marks.

Conclusions

The systematic mapping of cross-bedding and other primary current structures contributes materially to reconstruction of the conditions of deposition and paleogeography of ancient sand rocks. The cross-bedding of the Precambrian quartzites studied has a marked preferred orientation, and this orientation is uniform over a very large region and through a long period of time (great thickness of strata). These observations suggest one or more periods of accumulation of quartz sand derived from the Temiskaming and Keewatin provinces of the Shield north and northwest of the present quartzite outcrops.

The inclination of the cross-laminations commonly exceeds the angle of repose, and such aberrant dips are ascribed to deformation during the tilting of the beds, as the average inclination in a given area is closely correlated with the mean dip of the beds in that area.

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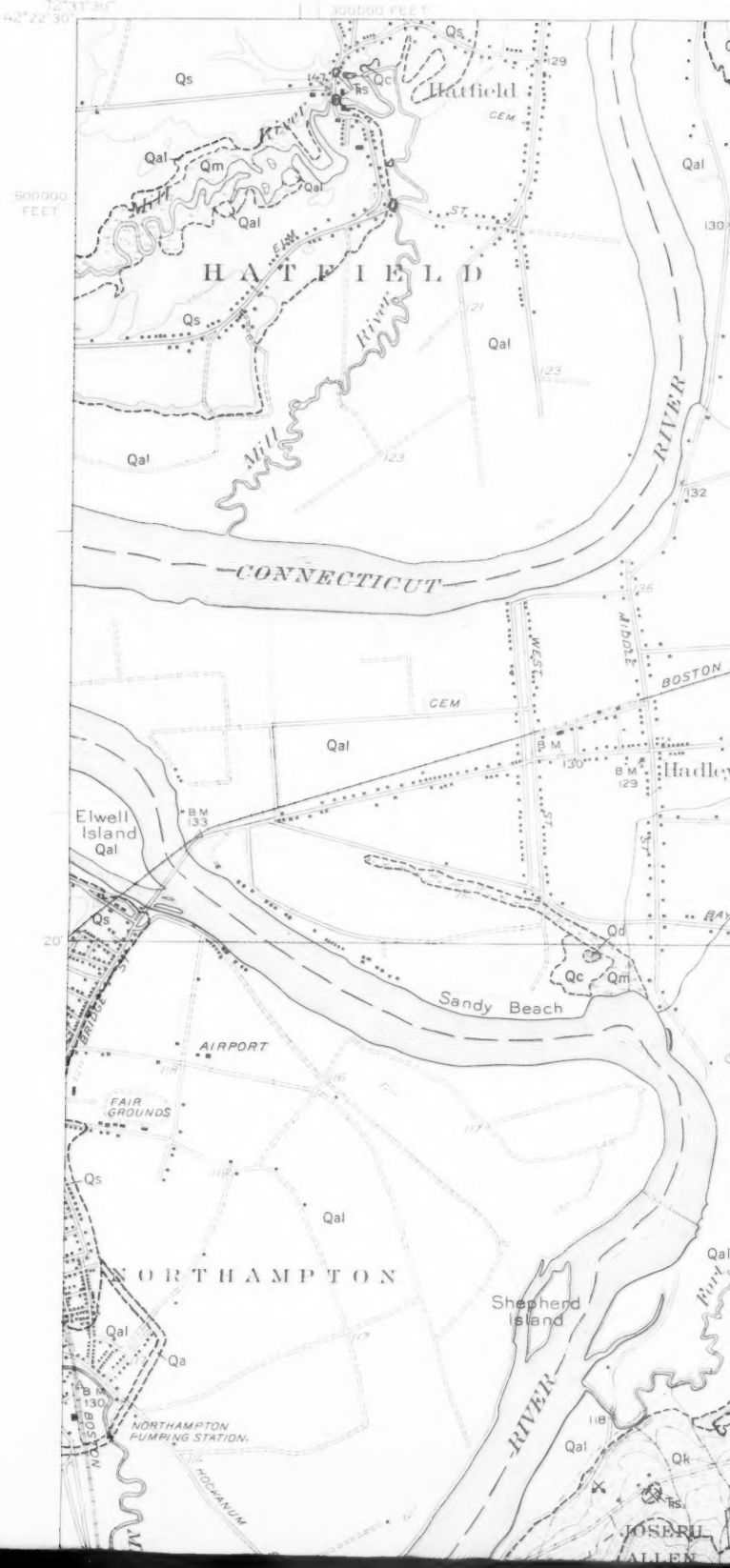
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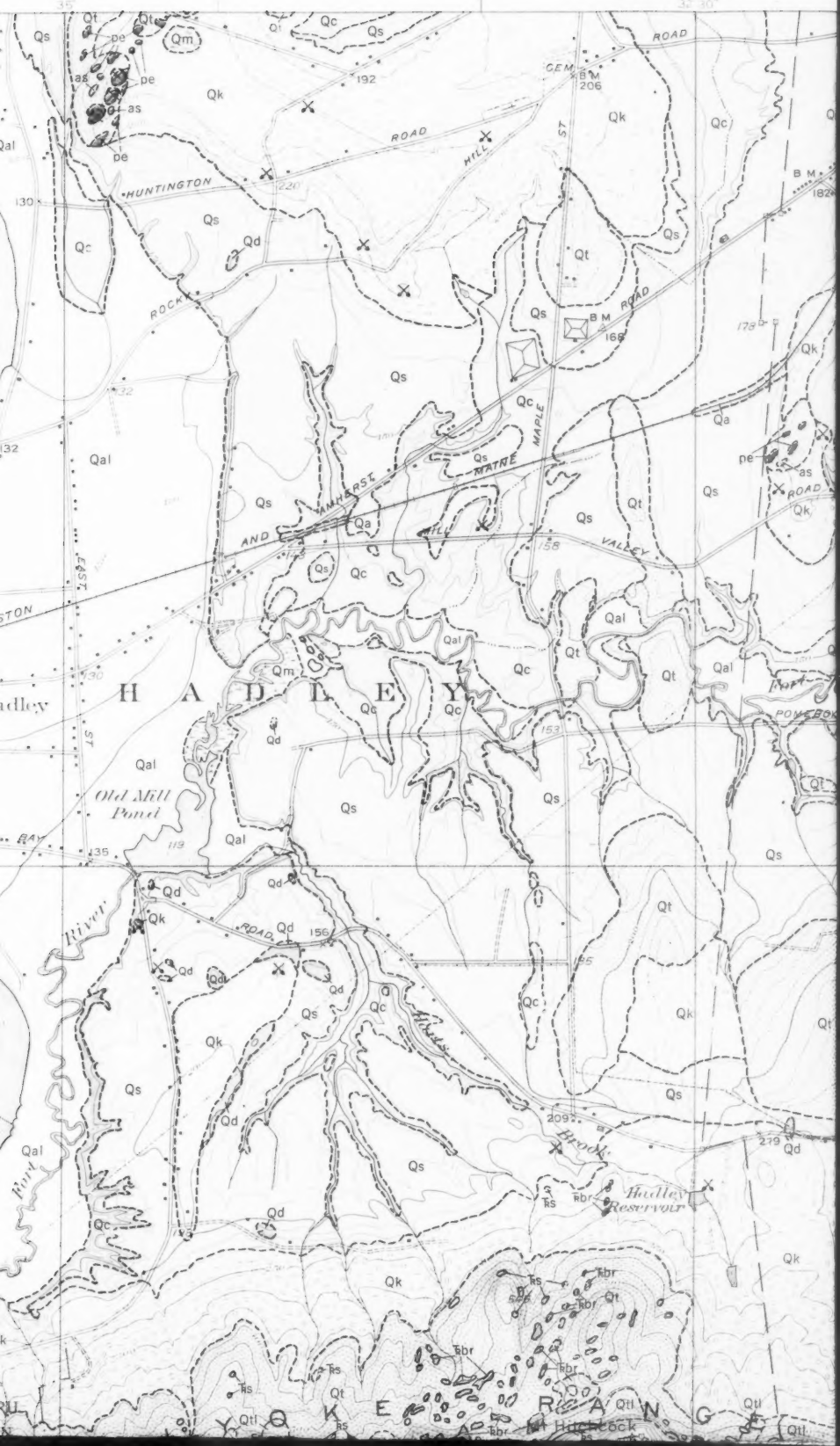
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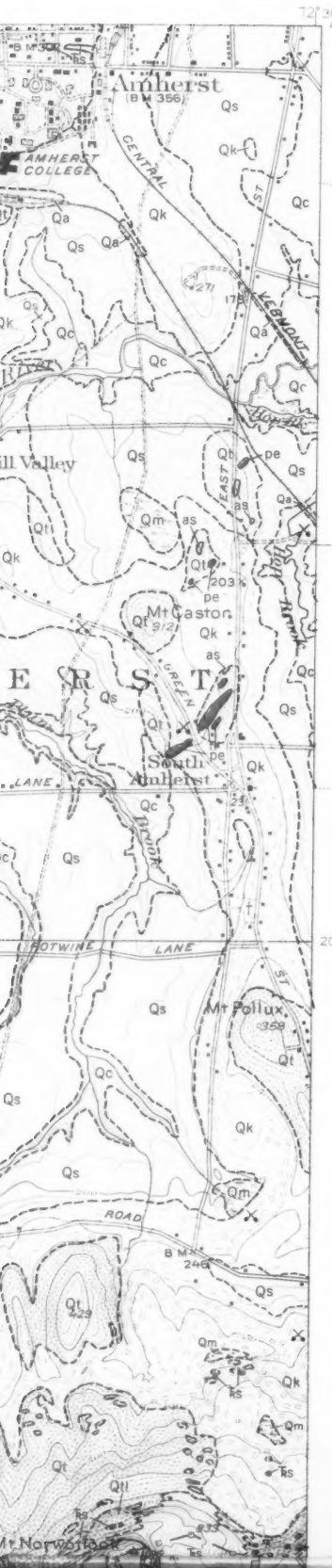
POWDER
SOUTHEAST
BAY
MAPLE
192
206
182
166
132
130
129
135
156
209
279
153
63



EXPLANATION

QUATERNARY

Post-Pleistocene

Qa — **Artificial fill**

Qm — **Peaty clay and muck**

Qtl — **Talus**

Qd — **Sand dune**

Qal — **Alluvium**
Flood-plain deposits of silty clay, silt and fine sand

Pleistocene

Qs — **Fine sand and silt**
Gray to pale yellow, passing into coarser sand and gravel and (basinward) into silty clay

Qc — **Varved clay**
Light or dark gray, locally purplish, seasonally stratified clay

Qk — **Kame gravel**
Compact, or indistinctly stratified, gray or yellowish coarse gravel, with cobbles and pebbles derived from till, passing in places into till

Qe — **Esker**
Gray or pale yellow-brown, cross-bedded gravel, with lenses of sand and finer gravel; forms elongate low hills

Qt — **Till**
Gray, locally purplish, compact aggregate of boulders, cobbles, and small fragments in groundmass of clay, silt, and sand

Triassic

Trbr — **Black Rock diabase breccia**
Fragments of diabase, tuffaceous diabase, tuff, and sedimentary rocks in matrix of vesicular or compact diabase

32' 30"
Hatfield
Hadley
H A D L E Y
Old Mill Pond
Sandy Beach
Shepherd Island
RIVER
Hadley Reservoir
Brook
ROAD
BOSTON
ROCK
MAPLE
HUNTINGTON
AND
AMHERST
VALLEY
POWDER
MILL
WEST
EAST
MIDDLE
SOUTHEAST
CEM
BM 206
BM 182
BM 130
BM 129
BM 168
Qs
Qt
Qc
Qal
Qd
Qk
Qm
Qa
Qbr
Ts
Qu

EXPLANATION

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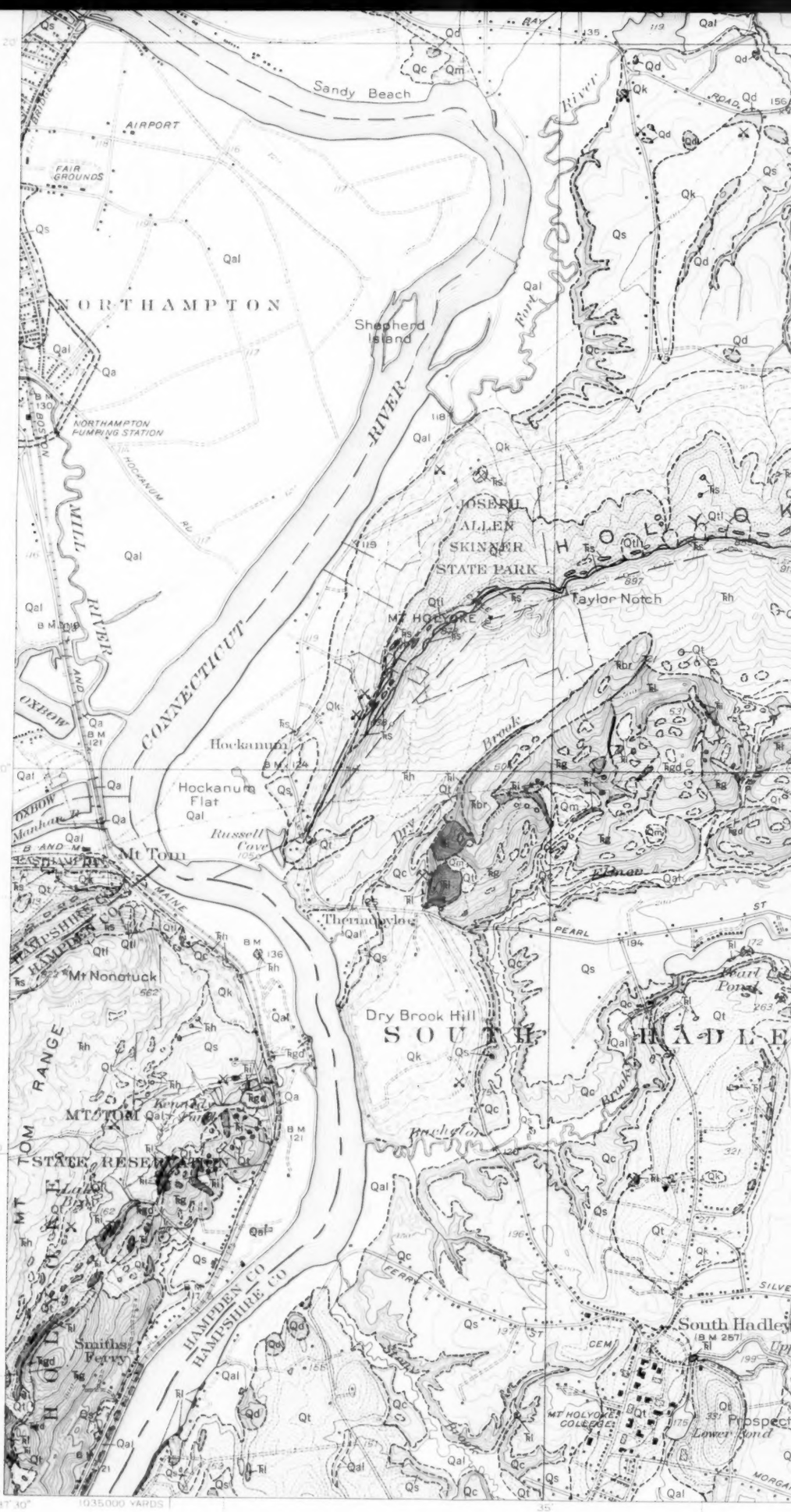


GEOLOGIC MAP OF MOUNT HOLYO

APPROXIMATE MEAN DECLINATION, 1947

MAGNETIC NORTH
TRUE NORTH

1000 0 1000 2000

CONTOUR
DATUM

32°30'

320000 FEET

LYOKE QUADRANGLE, MASSACHUSETTS

SCALE 1:31680

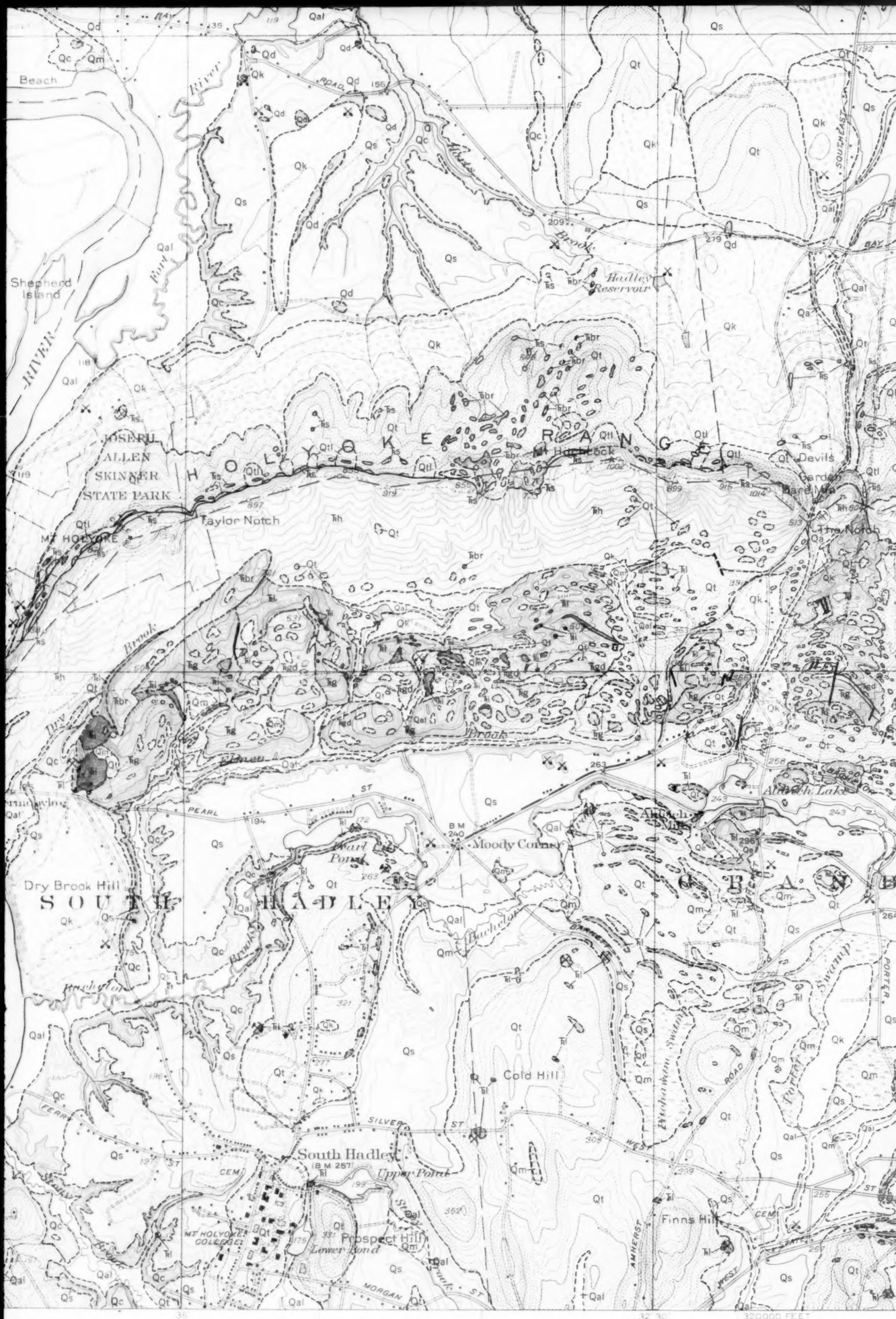
0 1 MILE

0 2000 3000 4000 5000 6000 7000 FEET

0 1 KILOMETER

ONTOUR INTERVAL 10 FEET

DATUM IS MEAN SEA LEVEL



GIC MAP OF MOUNT HOLYOKE QUADRANGLE, MASSACHUSET

SCALE 1:31680

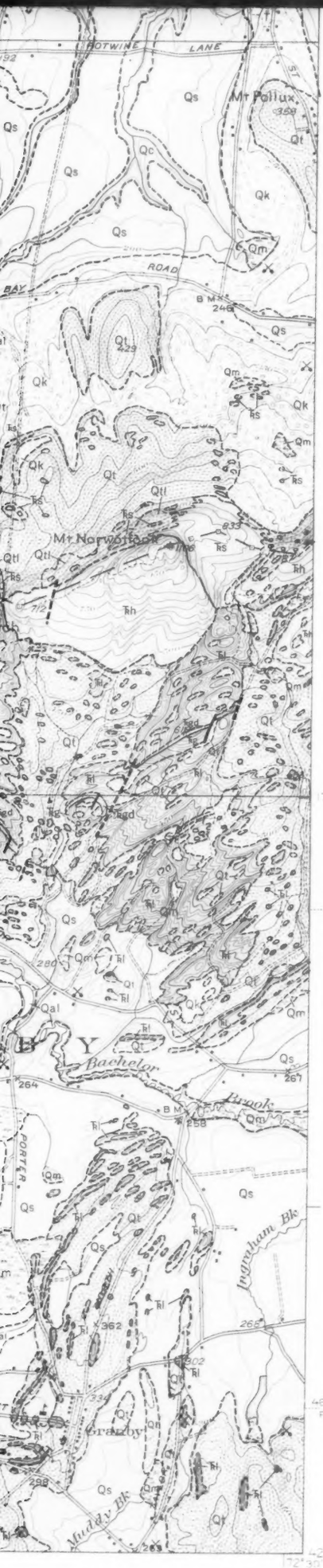
1 MILE

1000 0 1000 2000 3000 4000 5000 6000 7000 FEET

1 .5 0 1 KILOMETER

CONTOUR INTERVAL 10 FEET

DATUM IS MEAN SEA LEVEL



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Granby tuff

Trbr

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Fragments of diabase, tuffaceous diabase, tuff, and sedimentary rocks in matrix of vesicular or compact diabase

Tri

Intrusive diabase pipes, dikes, and sills of fine-grained diabase

Trgd

Diabase flow in Granby tuff

Fine-grained, locally altered, compact or vesicular diabase

Trg

Granby tuff

Brown, purple, or maroon, compact or stratified diabase tuff, or tuffaceous sandstone

Trl

Longmeadow sandstone

Brick-red, purple, or yellowish-gray, cross-bedded feldspathic sandstone and conglomerate, in upper part finer-grained, with siltstone lenses. Lower part older than Granby tuff, upper part later than tuff

Trh

Holyoke diabase

Greenish-gray, compact, fine- to medium-grained, locally vesicular diabase

Trs

Sugarloaf arkose

Purplish, brick-red, or yellowish medium- to coarse-grained, cross-bedded arkose with conglomerate lenses

TRIASSIC

pe

Pegmatite lens

Coarsely crystalline aggregate of quartz, oligoclase, potash feldspar, and subordinate mica

as

Amherst schist

Dark gray or brownish foliated muscovite-quartz schist, with subordinate feldspar and biotite

PRE-TRIASSIC

Contact, dashed where approximately located

Fault, dashed where inferred

Dinosaur tracks

Quarry

Sand or gravel pit

GIC MAP OF MOUNT HOLYOKE QUADRANGLE, MASSACHUSET

SCALE 1:31680

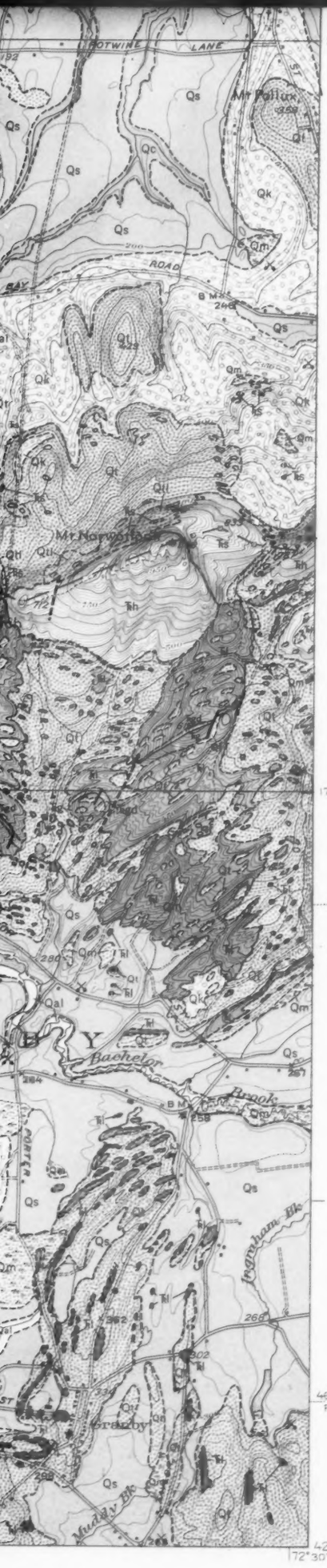
1 0 1 MILE

1000 0 1000 2000 3000 4000 5000 6000 7000 FEET

1 .5 0 1 KILOMETER

CONTOUR INTERVAL 10 FEET

DATUM IS MEAN SEA LEVEL



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Coarsely crystalline aggregate of quartz, oligoclase, potash feldspar, and subordinate mica

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Dark gray or brownish foliated muscovite-quartz schist, with subordinate feldspar and biotite

PRE-TRIASSIC

Contact, dashed where approximately located

Fault, dashed where inferred

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 68, PP. 481–504, 6 FIGS., 3 PLS. APRIL 1957

GEOLOGY OF MOUNT HOLYOKE QUADRANGLE, MASSACHUSETTS

BY ROBERT BALK

Preface

Robert Balk was killed on February 19, 1955, in the tragic plane wreck on Sandia Mountain near Albuquerque, New Mexico. The manuscript on the Mount Holyoke quadrangle had been submitted, and the Society has undertaken its publication as a memorial to its author.

In submitting the manuscript to the editor Robert Balk wrote " . . . as a profession we must, I think, also consider means and ways of introducing our science to students in a reasonably up-to-date and reliable manner in such a way that they find features described and discussed which they can see and study at their leisure *while they are students*. I have therefore a modest hope that the map will be of some real value to the many hundreds of geology students who are being educated year after year, in the Connecticut Valley in sight of Mt. Holyoke."

Balk was stimulated to map Mount Holyoke quadrangle by an overdose of indoor work largely due to a very heavy teaching load during the war. He taught over thirty hours a week and needed field work.

On January 17, 1944, he wrote: "I just found out that we will have about three weeks of spring vacation in May and I have decided to map a large portion of the Holyoke quadrangle. The Holyoke Range, with its somewhat complicated basalt flows and tuffs, is almost done and what remains is quite simple: silt terraces with clay above Triassic sandstones and shale. I have been invited to stay on a farm in the northwestern part of the quadrangle as headquarters. If this works out I could complete field work by fall and then write a quadrangle report, privately mapped and financed and send it with a map to G. S. A."

Toward the end of February he had been in the field during several weekends and wrote on February 26, "After three weeks the snow is finally gone and I had some good days in the quadrangle. Two more months before it will be hot and I will complete the Quaternary soon. I found some huge erosion gullies which trouble the tobacco and onion farmers quite a bit. But it is interesting to see how fast they grew. Some have increased one-third during the last eight years since the topo maps have been made. They are all cut into varved clay. The gullies were started where animals had made trails."

That summer of 1944 proved to be disappointing because much time had to be spent on summer-school teaching and there were only weekends for work on the quadrangle. He admits his disappointment and says, "We are spring chickens no longer and the years run by like thin sections which one looks at to see what one can make of them, but already something else is waiting and needs to be done. Such is life!"

Finally in 1946 he had completed the field work, and only the report and final drafting remained. But new projects and duties appeared, and completion has been long delayed.

The Holyoke quadrangle was done without the use of a car, just as previous work in the Clove quadrangle and the Adirondacks. He went to any length to reach important outcrops and paid no attention to personal comfort.

In the summer of 1946 he describes his working day as follows: "For two weeks I am now in the Adirondacks—unfortunately still without a car, but I have to overcome terrible distances. Breakfast not before 7:30. Then two and one-half hours to a farm where I reach the cowtrail through a swamp and it takes about forty minutes through the brush to get to the exposures. Then I have about three or four hours time for some geology and after that there are three hours

to walk back to the Inn. The difference in elevation is 2500 feet! But I think it is worth it. The other day I met with an augite crystal ninety-one by thirty-eight centimeters and I lifted my hat."

Robert Balk was dedicated to geology and field work as few other men were. He was happiest when he could *see* geology and study the rocks where they grow. His longing for field work was in irreconcilable conflict with his keen sense of duty. He was paid by the college to teach and he taught a large number of courses, especially during the war. And he was a superb teacher who could talk about his subject easily and from an excellent memory. Many of his students who remember little else of their college days remember Balk's lectures. To hear him talk geology was a rare treat, and his audiences were fascinated by his presentations.

He rarely rested or relaxed in the way others do. When urged by his friends to rest and take a vacation he would answer, "If I want to really relax I can do that best and most agreeably with an interesting, undisturbed, and unhurried piece of work for which I can take my time and where my thoughts are not constantly forced in other directions. If I do nothing or play—I become terribly nervous." The conflict is common wherever undisturbed research conflicts with heavy teaching loads and hardly unique to Holyoke or Balk. But the solution of the conflict is that of a dedicated man with enormous energy: he doubles his working time and gives up his personal comfort. Thus he has published many outstanding papers and made his lasting contribution to the science of geology, in spite of the fact that he had no time for research. The urge of the devoted scientist was paralleled by an above average sense for civic responsibility. Balk keenly felt the lethargy and general lack of interest in public affairs, especially during the war. It was hard for him to understand how anyone could live in the richest and most prosperous land on earth and take so much for granted. Paying taxes was a pleasant duty for him, and he was seriously disturbed about the fact that people will buy cars and many luxuries but only hesitatingly would pay taxes to educate their children.

The Mt. Holyoke quadrangle is not an ordinary paper on an average region but is meant by the author to be used by students who want to go into the field and study the area on their own. It is meant to help the younger generation to *see* geology and to observe without the guidance of older and more experienced men. The paper contains much new and original data but is written as an educational description to a map in an area where three colleges train hundreds of young men and women in geology, according to Balk, "practically within walking distance."

ERNST CLOOS

ABSTRACT

The oldest bedrock formations in the Mt. Holyoke quadrangle, Massachusetts, are feldspathic schists, tentatively correlated with Emerson's Amherst schist, and pegmatite lenses, perhaps of late Paleozoic age. Next in order of decreasing age are the Triassic Sugarloaf arkose and Longmeadow sandstone, with a 300- to 400-foot diabase flow between them. The Granby tuff, with associated lava, pipes, sills, and dikes of diabase was erupted in early Longmeadow time and records an unusual phase of Appalachian Triassic vulcanism.

The bedrock floor is extensively covered by unconsolidated Pleistocene and post-Pleistocene deposits. Only one till, presumably Wisconsin, is recognized. In the Connecticut Valley proper, lacustrine and outwash deposits of varved clay, silt, sand, and gravel are widespread. Recent flood-plain deposits consist of silt and sand.

Rural settlements and dairy farms derive nearly all the water from the top of the late Pleistocene varved clay, an ideal aquifer at shallow depth. The Triassic sedimentary rocks could supply copious amounts of water, and artesian conditions should exist over part of the area.

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Introduction

Mount Holyoke quadrangle, about 13 miles north of Springfield, Massachusetts, includes a scenic area in the Connecticut Valley where the Holyoke Range, a prominent trap ridge, abandons its normal north-south course and veers to the east (Fig. 1). Where it curves, the ridge is breached by the Connecticut River in a picturesque gorge, the "Narrows," that separates a wide northern lowland from a narrower valley on the south in which Triassic bedrock is exposed.

These and other geologic features of interest caused the writer to map the area, but the chief reason was educational. Within the quadrangle, or a few miles outside of it, are located the University of Massachusetts and Amherst College at Amherst, Smith College at Northampton, and Mount Holyoke College at South Hadley. All these institutions have active geological departments. The writer regards it as desirable that geology students have at their disposal a reasonably accurate geological map at the place where they receive their first geological instruction. Earlier geological maps of this part of Massachusetts (Emerson, 1898a; 1898b; 1917) are on small scales; a new set of topographic maps, published by the United States Geological Survey during the 1930's, on the scale of 2 inches = 1 mile, facilitated geologic mapping of the region. The field work was done in 1942–1946 when the writer resided at South Hadley.

Acknowledgments

The writer is grateful to The Geological Society of America for providing the funds for four of five new chemical analyses. Thanks are due Dr. John C. Haff for lending a number of thin sections from the petrographic collections of Mount Holyoke College. For valuable discussions of the mode of origin of the late Pleistocene deposits the writer is indebted to Richard H. Jahns and Max E. Willard. Residents of Amherst, Hatfield, Hadley, and South Hadley, too numerous to mention, gave the writer pertinent data on well records and groundwater conditions in the area.

Scope of Study

Apart from preparing an accurate map, the writer tended to supplement Emerson's voluminous descriptions of the region (1898a; 1905; 1917) with observations on the various phases of diabase in this district, and give descriptions of exposed structural features in the Triassic rocks. The disastrous flood of the Connecticut Valley, in March 1936, made it clear that detailed information on the Quaternary deposits is of paramount importance here. Facts and figures on these rocks aim to supplement Jahns' study of the flood-plain deposits in the Connecticut Valley (1947). Geologists, whether teachers or students, are urged to note and record errors and omissions on the map, so that eventually a truly accurate map may be available.

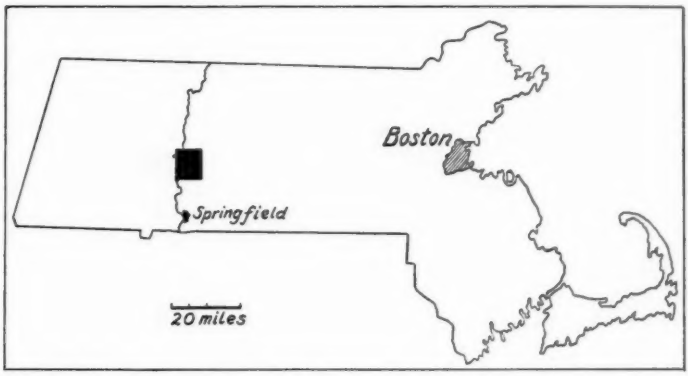


Figure 1.—Index Map of Massachusetts, Showing Location of Mount Holyoke Quadrangle

General Geology

The curving trap ridge, known as the Holyoke Range on the east side of the Connecticut River and Mount Tom Range on the west, provides, with its steeply north-, northwest-, and west-facing bold cliffs, the most striking scenic feature of the quadrangle. It also sharply outlines the northwestern border of a Triassic sedimentary basin that is floored over a considerable distance by the trap sheet. The center of the basin—referred to as the southern one, in contrast with a northern one between Amherst-Hatfield and Bernardston-Northfield (Bain, 1932, Fig. 1)—lies south of Mount Holyoke quadrangle.

North of the Holyoke-Mount Tom Range, the few small outcrops of consolidated bedrock are apparently the deeply eroded remnants of a broad swell that separates, as it might have in Triassic time, the northern and southern Triassic basins. Even pre-Triassic rocks appear at the surface. The area also furnishes a fairly complete section through the glacial and postglacial deposits of this part of the Connecticut Valley. The evolution of the modern flood plain of the Connecticut River from late-glacial conditions can be reconstructed here with a fair degree of accuracy and is substantially in accord with the results of Jahns' and Willard's work (1942).

Pre-Triassic Crystalline Rocks

Mount Warner stands 1 mile east of the Connecticut River, and just north of Mount Holyoke quadrangle. It is underlain by feldspathic schists (Willard, 1952a), and outcrops on its south slope are the largest exposures of these ancient rocks in the quadrangle. Smaller ledges crop out 1½ miles southwest of Amherst and at scattered points near South Amherst. In several old test pits on the south slope of Mount Warner a dark-gray to brown, well-foliated, medium-grained schist is exposed, dipping at gentle angles northwest, and penetrated by several concordant and discordant pegmatite lenses. The folia of muscovite and

biotite (pleochroism light brown to tan) are dusted with magnetite. Between them lie flattish grains of quartz, oligoclase (an_{16}), a little orthoclase, microperthite, and a few grains of garnet and tourmaline. Schist at the other localities shows essentially the same composition, but at South Amherst were found lenses of a greenish-gray calcareous schist, composed of quartz, bytownite (an_{76}), clusters of chlorite enclosing remnants of a weakly pleochroic amphibole, sphene, pale-pink to colorless spongiform garnets, and magnetite. Emerson (1917, p. 75) called these rocks the Amherst schist and assumed Carboniferous age (1917, table opposite p. 17). Willard (1952a) considers the rocks Paleozoic (?).

The pegmatite lenses that cut the schist are composed in order of decreasing abundance, of quartz, oligoclase, orthoclase and microcline, muscovite, a little magnetite, and altered biotite. A few lenses contain small grains of red garnet and black tourmaline and some visible graphic intergrowth of quartz and microcline. A pegmatite lens on the south slope of Mount Warner seems to lack potassium feldspar, and some phases pass into medium-grained granite.

According to Bain (1941, Fig. 2), Triassic sediment occurs between the Amherst and Mount Warner areas of crystalline rocks. The writer was unable to locate the outcrops or obtain records of wells that penetrated Triassic rocks. The volume and thickness of these must be very small, and the northeastern portion of the quadrangle can be considered a block of crystalline rocks covered by a thin veneer of Quaternary deposits.

Triassic System

General Statement

The remaining bedrock formations in the quadrangle are Triassic. Four subdivisions are recognized, from bottom to top: (1) the Sugarloaf arkose (Emerson, 1899a, p. 354–358); (2) the Holyoke diabase (Emerson, 1898a, p. 446–481); (3) the Longmeadow sandstone (Emerson, 1898a, p. 346–369), a series of sandstone, conglomerate, and siltstone beds, becoming finer-grained near the top; (4) a series of tuffs, with interbedded tuffaceous sandstone and conglomerate, diabase, and intrusive pipes, dikes, and sills, referred to as the Granby tuff (Emerson, 1898a, p. 476–481). A fifth formation, the Chicopee shale (Emerson, 1898a, p. 370–371) overlying the Longmeadow sandstone, has not been observed within Mount Holyoke quadrangle, but thin, scattered lenses of silty shale were noted in a few places near the southern border of the mapped area.

All stratigraphic units participate in the synclinal structure of the southern basin of Triassic rocks. However, the Sugarloaf arkose also extends as a widespread blanket of irregular distribution and thickness northward over the crystalline rocks, and farther north forms the floor of the northern Triassic basin. The two basins show an almost identical sequence of rocks, but individual lithologic units have not yet been correlated.

Sugarloaf Arkose

Sugarloaf arkose crops out in three separate areas that may or may not be connected beneath the Quaternary deposits. Arkose at Hatfield, on the west side of the Connecticut River, has been traced by Willard (1952a) for 2 miles northward, where it continues northeastward across the river to North Hadley, at the west base of Mount Warner. A few small outcrops in the town of Amherst are believed to form a thin, discontinuous veneer over the Amherst schist. Mr. James Tufts informed the writer that red sandstone was found in an excavation at the road intersection at the north-northwest base of Mount Pollux, and the cuttings of a well on the property of Mr. N. James Schoonmaker, at the west base of the same hill, appeared to be Triassic sediment, but, as the material had lain on the ground for some time before it was examined, the identification is not conclusive.

The third outcrop zone is a long, narrow belt, extending from the eastern quadrangle border westward along the north slope of the Holyoke Range. Southwest of Mount Holyoke it crosses the river and continues southwestward at the northern base of the Mount Nonotuck-Mount Tom Range beyond the western border of the quadrangle.

The Sugarloaf arkose is coarse- to medium-grained and is brick red, purplish, or yellowish gray. Films and thin discontinuous lenses of finer-grained siltstone separate the coarser

phases into beds 1–6 feet thick. Pebbles in coarse layers are, in order of decreasing abundance, vein quartz, aphanitic quartzitic rocks, pegmatite, various gray, pink, and reddish gneissic granites, and some greenish-gray

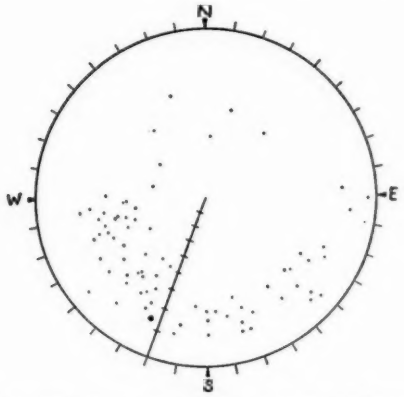


FIGURE 2.—DIRECTION AND ANGLE OF DIP OF FORESET BEDS IN SUGARLOAF ARKOSE AND LONGMEADOW SANDSTONE IN MOUNT HOLYOKE QUADRANGLE

Shown in stereographic projection (lower hemisphere). Key: heavy dot represents a bed that dips at 13° in direction S. 24° W. Cross bars on radius represent (from center outward) intercepts of arcs denoting 80°, 70°, 60°, *etc.* through 10-degree inclination. The majority of foreset beds dip W.-SW. at about 20°–30°. Field measurements have been rotated so that the associated bedding planes are horizontal.

siliceous phyllites. Though the innumerable quartz veins of the Berkshire Hills west of the Connecticut Valley could have supplied many quartz pebbles, the abundant feldspar debris in the Sugarloaf arkose suggests that the main source lay east of the present valley. This is in agreement with observations of Bain (1941, p. 267) in Massachusetts and of Longwel (1922, p. 232) and Wheeler (1937, p. 66, 67) in Connecticut. It is also supported by the west-southwesterly dips of the foreset beds in cross-bedded sections (Fig. 2). In general, the grain size decreases from east to west (Pl. 2). However, outcrops of the arkose in this quadrangle are so small that only a few attitudes could be observed. The arkose has calcareous cement with or without iron oxide.

The thickness of the formation cannot be determined in this quadrangle because of lack of continuous exposures. Bain (1941, p. 274, 275) suggests 3000–6800 feet. As the arkose overlies a highly irregular floor of older rocks, the variations in thickness are understandable. (*See* Willard, 1941; 1952b.)

Holyoke Diabase

The diabase sheet that underlies the Holyoke and Mount Tom ranges is the northernmost part of a north-south trending ridge that can be traced southward at least to West Hartford, Connecticut, a distance of about 35 miles. It probably continues to the New Haven district. Although the entire zone of curvature from the regional north-south direction to east-west is within the Mount Holyoke quadrangle, the Holyoke Range continues for about 1½ miles beyond the eastern border of the map, where it is broken by several faults, loses elevation, and does not reach the eastern border fault of the Triassic fault block (Bain, 1941). In the Mount Holyoke quadrangle the diabase sheet is 300–400 feet thick. As it rests on much less resistant rocks, the diabase forms continuous, bold cliffs near the crest and along the outer (northern, northwestern, and western) slopes of the curving ridge. On the inner side, it supports smooth, monotonous dip slopes, almost free from till cover (Fig. 3).

The diabase is a fine-grained, massive, dark-greenish-gray rock, showing sparse white-weathering feldspar phenocrysts, 1–2 mm across. On fresh surfaces only innumerable small reflecting surfaces of feldspar grains are visible. Pyroxenes are difficult to identify, and most magnetite grains are too small to be recognized. Weathered surfaces are lighter shades of greenish gray, even white, and advanced chloritization of pyroxenes produces patches of soft green. Most vesicular or hydrothermally altered diabase is pale green or mottled with rusty colors.

The diabase does not exhibit layers of contrasting texture or mineral composition and apparently is one flow. However, in all larger exposures the diabase shows smooth fractures, parallel to the floor. They appear at all levels but none could be traced continuously from one ledge to another, which would be possible if they represented boundaries of individual flows (Fig. 4). Most of the surfaces are coated

with a thin light-gray or white film, a powder of which has the optical properties of quartz or chalcedony, and is light greenish where minute scales of chlorite are present. The fractures were not seen in vesicular diabase. Possibly they represent subparallel planes along which entrapped gas and vapor were squeezed out of the cooling lava. The experiments of Birch (1933) on the influence of escaping gas on the layer structure of clays may be significant in this connection. A mass of soft clay that contains gases and is being compressed develops separation planes parallel to the compressing wall, along which the enclosed gases escape. Soft, gas-charged lava near the floor of a thick flow would present a similar situation, and the veneers of silica and chlorite could be interpreted as some of the mineral matter precipitated by the escaping hot vapors. Under one set of conditions the volatiles were squeezed out, and the fractures formed; under slightly different conditions the gases remained in the surrounding rock and formed vesicles.

Columnar jointing is well developed in the west- and north-facing cliffs of the range. At several places, especially close to the arkose floor, the columns are dissected by downward-converging fractures, so that some overhanging cliffs whose floor rocks have crumbled look like rows of immense projectiles (Fig. 5). The reason for these peculiar joints is obscure, but they are not confined to the Holyoke-Mount Tom Range. Lewis (1908, Pl. 37, fig. 2) describes and illustrates a "sphenoidal jointing" in the Palisades diabase. He also mentions another peculiar and common type of fracture in columnar diabase in which cracks across the column axes are not straight but curved like shallow saucers. Lewis (1908, Pl. 37, fig. 1) speaks of "ball-and-socket" joints, and they were seen in several cliffs of the Holyoke Range. Almost everywhere the fractures are concave upward, but the opposite was also observed. In contrast to the larger fractures parallel to the floor, the crack systems are obviously due to special cooling or shrinking stresses in each column.

Laths of subhedral plagioclase, seriate from 0.1 mm to 2 mm in length, form a mesh in the interstices of which smaller equant anhedra of almost colorless clinopyroxene are embedded. Blebs of partly devitrified pale-brown glass

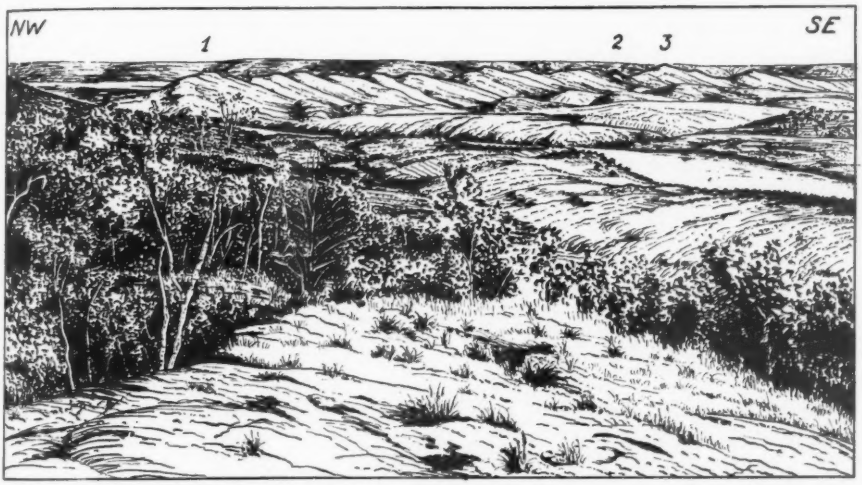


FIGURE 3.—VIEW NORTHEAST FROM GOATS HEAD MOUNTAIN, SOUTHWEST OF MOUNT NONOTUCK

Holyoke Range in middle distance, with conspicuous dip slope to the south. (1) Mount Holyoke; (2) the Notch; (3) Mount Norwottock. Flat skyline in distance is Worcester upland of crystalline rocks. Ledge in foreground is Holyoke diabase dipping east, into the southern Triassic basin. Low hills at base of slope are Granby tuff; on far side of the Connecticut River, near center of view, Dry Brook Hill; beyond it, at south base of Holyoke Range, Black Rock and, farther off, hills of Longmeadow sandstone and Granby tuff

with minute needles of albite, orthoclase, and a trace of quartz, magnetite anhedra, and a few small prisms of apatite complete the simple

Clinopyroxene is pigeonite, with α = 1.688-1.695, β = 1.697–1.701, γ = 1.722–1.725 (specimens from the Notch quarry, and Mount

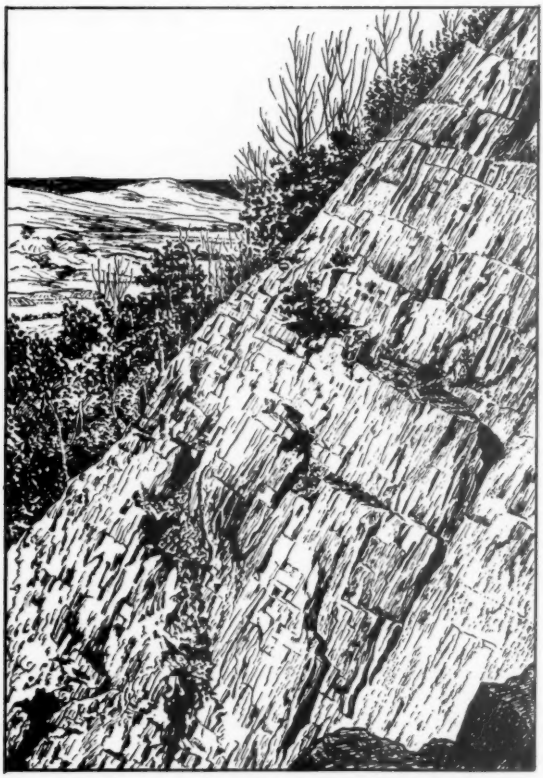


FIGURE 4.—LEDGE OF HOLYOKE DIABASE ON NORTHWEST SLOPE OF MOUNT NONOTUCK
Showing crude columnar structure and numerous smooth fractures parallel to the floor of the trap sheet.

mineralogical composition of the rock. Olivine, reported from one locality (Emerson, 1898a, p. 493), has not been found despite a special search; orthopyroxene, likewise alleged to be present (Stevens, 1938, p. 132), was sought in vain.

In 16 thin sections and powder preparations the plagioclase ranges from labradorite, An_{52}, to An_{69}. Individual, normally zoned crystals gave compositions of An_{61-54}, and in a few coarse, pegmatitic veins the plagioclase varies from An_{48} to An_{56}. Variations between cores and margins of plagioclase grains rarely exceed 6 per cent An.

Norwottock). In 22 measured grains from six localities 2v(+) = 39°–62°, with the greatest number around 40°–50°. Only three grains were found with 2V = 14°, 4°, and 0°. In earlier studies of the trap rocks of the Greenfield-Northfield district to the north, somewhat more clinopyroxene with small axial angles was found. Zoning is rare, but one grain with small axial angle has a shell of diopsidic pyroxene, with 2V(+) = 44°. Twinned pyroxene grains are abundant, but the minute lamellar twinning known from other diabase sheets was not seen.

Numbers 1 and 7 of Table 1 give chemical

analyses of the freshest diabase found. For comparison are given two analyses by Stevens (1938), made by semimicro technique (Nos. 3, 8), and an old analysis by Hawes (1875b) which until now has been the only chemical analysis of diabase from this area (No. 2). The new analyses agree closely with one from the middle portion of the Palisade diabase of New Jersey (No. 5), and also with one of intrusive diabase from Goose Creek, Virginia, well known through Shannon's careful study (No. 6). In general, the composition of the Holyoke diabase is close to the average values of Deccan and Karroo traps (Walker, 1940, p. 1081); clearly, it belongs to Kennedy's tholeiitic group (1933, p. 241).

Even in the freshest rocks some pyroxene has been partly chloritized. The alteration begins at the periphery, may consume entire grains, and is probably related to the growth of chlorite along the walls of minute vesicles that are found in every specimen of trap from this area. The most common chlorite is pale-green and weakly pleochroic. Rarely the mineral is colorless, and in many vesicles several shells of chlorite of contrasting color have been deposited. Among them are yellow and orange ones, but whether this denotes different

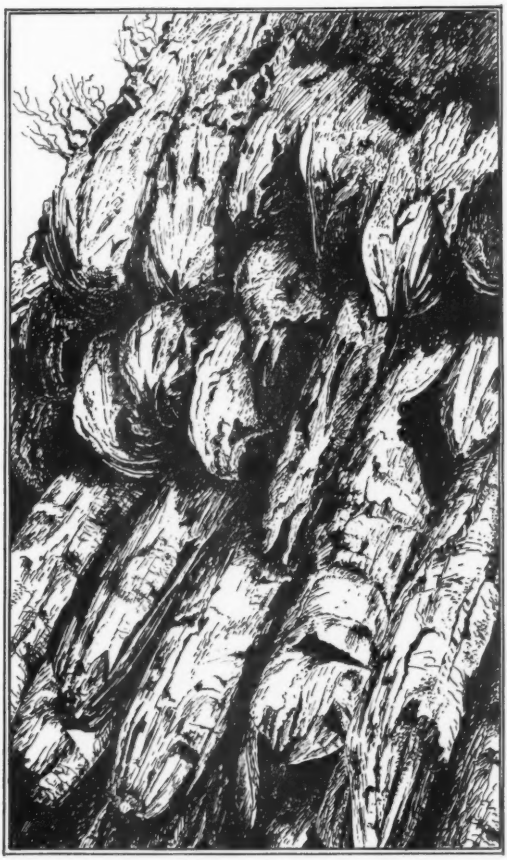


FIGURE 5.—COLUMNS OF DIABASE, TAPERING ALONG SECONDARY FRACTURES THAT CONVERGE DOWNWARD
Titan's Piazza, northwest base of Mount Holyoke

varieties of chlorite or is due to mechanically admixed iron oxide was not determined. The shells consist of minute fibers with axes normal to the outer walls of the vesicles. The interiors may be open or occupied by a few large calcite grains. Chlorite veneers on joints are common and closely resemble descriptions by Shannon (1924, p. 6) and Hawes (1875b). However, the material secured for tests was too impure and fine-grained to permit a check of whether or not this chlorite has the properties of diabantite.

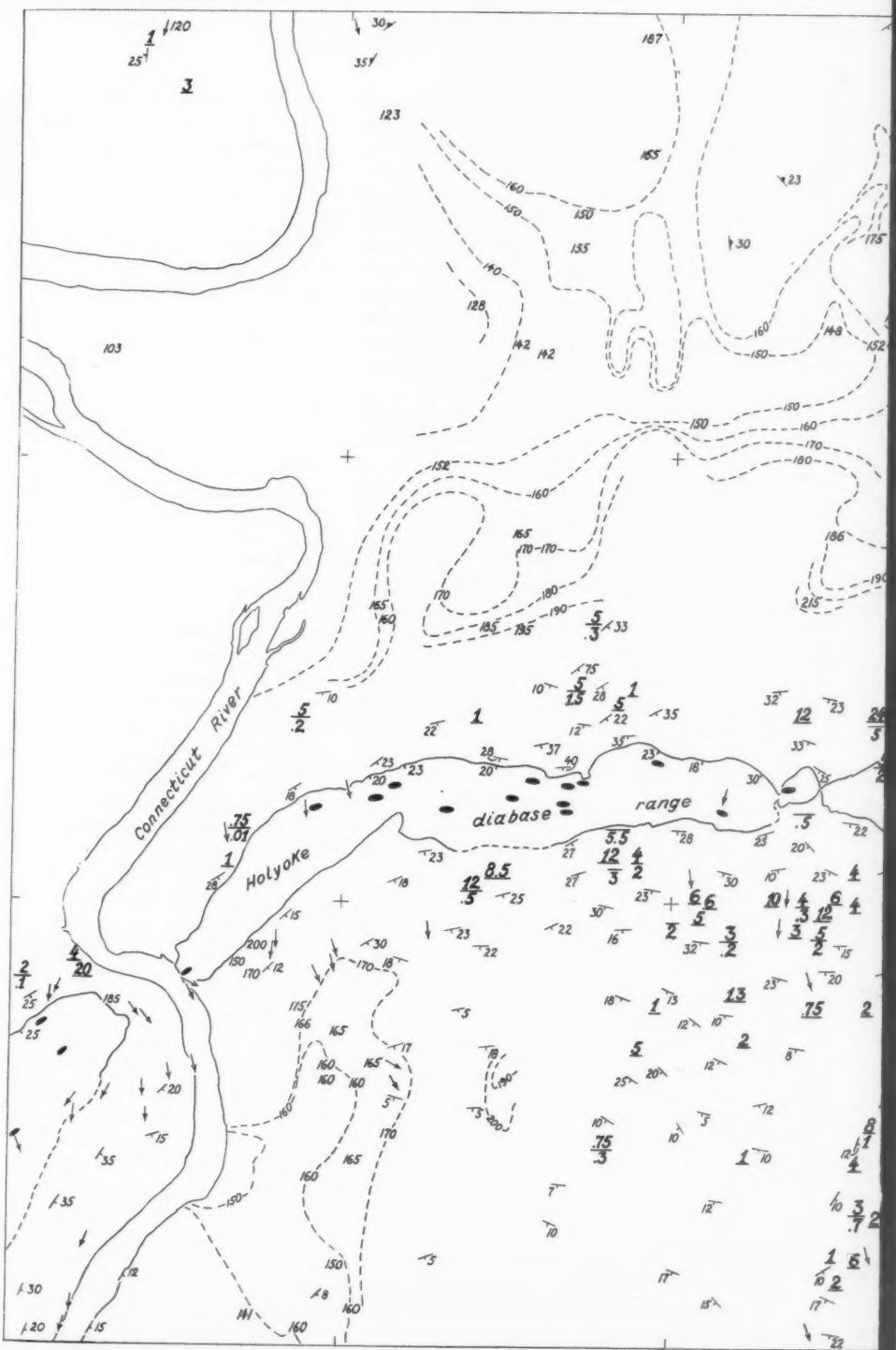
Plagioclase is less altered than pyroxene. Minute flakes of sericitic mica with or without associated carbonate are the common alteration products. In thin sections where as much as 10 per cent of the pyroxenes are chloritized, the feldspar may still be substantially unaltered.

Lenses of coarse, gabbroic diabase have been described from the Mount Tom diabase rang at Westfield, Massachusetts (Shannon, 1919) from southwest of Holyoke (Emerson, 190 1905), from the New Jersey Palisades diaba sheet (Walker, 1940, p. 1092–96), from the Goose Creek, Virginia, diabase (Shannon, 1924, p. 3), and from other places in the Appalachian Triassic belt. Similar rocks have been found in the Karroo dolerites (Walker and Poldervaart, 1949) and the Whin Sill (Tomkeieff, 1929). A systematic search of the literature would probably disclose many more examples. Emerson (1898a, p. 451) mentions gabbroic diabase from Mount Holyoke. On Plate 2, 17 masses of this rock are located. The most easily accessible outcrop, during the writer's survey, was in the northwestern part of the large Notch quarry. Here a lenslike mass

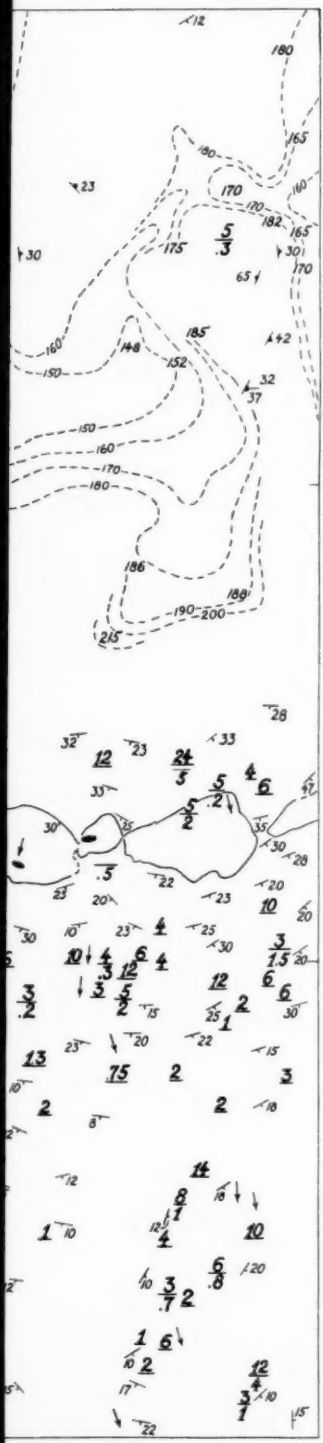
TABLE 1.—CHEMICAL ANALYSES OF FRESH AND ALTERED DIABASE

	1	2	3	4	5	6	7	8	9	10	11
SiO_2	51.58	52.70	50.41	51.78	52.32	51.56	49.49	50.25	49.17	43.92	43.08
TiO_2	1.13	n.d.	1.01	1.41	0.97	1.48	1.39	1.29	1.36	.85	.44
Al_2O_3	13.19	14.11	16.40	14.20	16.54	13.81	13.11	15.20	12.58	9.80	1.72
Fe_2O_3	3.54	1.87	.14	3.59	1.58	.96	3.23	.38	3.15	1.26	.16
FeO	9.66	9.78	10.33	8.25	8.66	11.32	11.43	12.59	11.36	8.80	4.90
MnO	.20	.45	.15	.43	.12	.19	.20	.18	.18	.18	.19
MgO	5.91	6.42	6.90	7.64	5.43	7.40	5.20	6.42	6.09	5.49	7.01
CaO	9.90	9.36	10.13	10.70	9.68	10.08	9.82	10.29	7.51	10.48	19.92
Na_2O	2.56	2.54	2.62	2.14	2.32	2.08	2.37	1.74	4.15	3.28	.23
K_2O	.74	.89	.82	.39	1.03	.96	.61	.57	.60	1.22	.35
P_2O_5	.12	n.d.	trace	.14	.06	.16	.17	0.00	.14	.01	.00
H_2O^+	1.23	n.d.	.92	.63	.84	----	1.00	1.03	3.19	3.55	2.23
H_2O^-	.08	n.d.	.46	n.d.	.40	----	.17	0.26	.11	.00	.07
CO_2	30	n.d.	n.d.	n.d.	.16	----	1.56	n.d.	.15	10.94	19.31
S					none						
ignition		1.61									
Total	100.14	99.73	100.29	99.89	100.11	100.00	99.75	100.20	99.74	99.78	99.61
sp.gr.	2.95			3.03	2.97		2.98		2.83	2.79	2.683

1. Fresh Holyoke diabase, crest of Holyoke Range, west base of Mount Norwottock. H. B. Wiik, analyst.
2. Holyoke diabase, Mount Holyoke, G. W. Hawes (1875a, p. 186).
3. Holyoke diabase, Notch quarry, N. P. Stevens (1938, p. 151); semimicro analysis.
4. Diabase, West Rock, New Haven, Connecticut, G. W. Hawes (1875a, p. 186).
5. Palisade diabase, about 500 feet above base, F. Walker (1940, p. 1080), F. Herdsman, analyst.
6. Normal intrusive diabase, Goose Creek, Virginia, E. V. Shannon, (1924, p. 13).
7. Fresh, intrusive diabase. Pipe, northeast of knob 490′, 1 mile southwest of the Notch, H. B Wiik analyst.
8. Black Rock diabase, N. P. Stevens (1938, p. 151); semimicro analysis.
9. Altered diabase, flow in Granby tuff, Hill 510′, east of power transmission line, south of the Notch, H. B. Wiik, analyst.
10. Hydrothermally altered diabase breccia. East base of hill 856′, crest of Holyoke Range. H. B. Wiik, nalyst.
a 11. Aphanitic quartz-carbonate rock, derived from diabase through hydrothermal alteration. Northeast ase of Mount Norwottock, H. B. Wiik, analyst.
b



MISCELLANEOUS NUMERICAL DATA, MOUNT HOLYO
QUADRANGLE, MASSACHUSETTS



Legend

50 Strike and dip of foliation in crystalline rocks

26 30 Foliation, and bearing and plunge angle of crystalline rocks

12 Strike and dip of bedding in Triassic rocks

Bold-face figures denote grain sizes in Triassic rocks, e.g. $\frac{3}{.2}$: maximum diameter 3 inches, average diameter .2 inches

Black lenses in Holyoke diabase are gabbro masses

Direction of glacial striae

160 Approximate level of top of varved clays, 10-foot contour interval. Individual figures denote elevations at exposed contacts

Two meridians and 2 parallels divide the quadrangle into 9 rectangles. These coordinates are ticked in map borders, and crosses denote the 4 intersection points in the map

1 mile

UNT HOLYOKE
TS

of coarse, massive diabase, about 4 feet thick, and exposed along the strike more than 28 feet, lies in normal, fine-grained rock without sharp boundaries. In some places the boundary zone is irregular and is traversed by stringers, veins, and irregular pockets of aphanitic, light greenish-gray rocks that may or may not be associated with, or pass into, white quartz-calcite veins. Some of these have gabbro on one side and normal diabase on the other.

Most of the other gabbro masses in the quadrangle resemble in shape that of the Notch quarry and are flat lenses, rarely more than 3 feet thick, but they may have considerable horizontal length. None was seen directly at the floor of the diabase sheet, though a few appear only a few feet above it.

Microscopically, the gabbro lenses show almost the same mineral composition as the normal diabase. Clinopyroxenes, as much as 4 cm long, are commonly bent blades that are noticed at once on fresh surfaces, whereas the small pyroxene anhedra in ordinary diabase are invisible to the naked eye. Curved pyroxene blades have been noted by Shannon, Emerson, and Tomkeieff in gabbroic lenses in the previously mentioned areas. In the congealing coarse rock the pyroxenes appear to have grown faster, or at an earlier stage, so that the meshwork is of pyroxene rather than plagioclase. Several samples of coarse pyroxene gave $\alpha = 1.693$, $\beta = 1.697–1.703$, $\gamma = 1.718–1.728$; $Z\wedge c = 45°$, $2V\ (+) = 40°–50°$, rarely 12°–20°. Between the pyroxene prisms, the plagioclase ($\beta = 1.554–1.561$, An_{42-56}), too, forms large but nearly equant anhedra. In some gabbros, magnetite develops blades, like the pyroxenes, several centimeters in length. In the vicinity of the aphanitic veins, alteration is intense and leads to fine-grained aggregates of chlorite, carbonate, sodic plagioclase, and quartz.

A few outcrops give a hint as to how these coarse-rock types have originated. Where Easthampton Road crosses the base of the Mount Tom trap sheet, three-quarters of a mile south of the summit of Mount Tom (7½′ Mount Tom quadrangle), a number of sandstone fragments have been kneaded into the base of the diabase. In contact with these the diabase is coarse-grained. In several places the edges of the sandstone blocks are fringed by sheaves of pyroxene; elsewhere, they are surrounded by a pale-gray, aphanitic rock. From the sandstone fragments, a few veinlike stringers of coarse pyroxene and feldspar crystals penetrate as much as 3 feet of the normal diabase, and much of this coarse rock is porous and shows chlorite patches replacing some of the pyroxene. The outcrops suggest that some of the incorporated xenoliths were moist and probably still soft. The vapor generated when the slabs were heated probably entered the surrounding diabase and speeded up the growth of the normal silicates. Where excess vapor was present, the newly formed pyroxenes and plagioclases seem to have been attacked by hydrothermal action and altered to chlorite and other secondary minerals. Emerson (1905, p. 117, 118), Shannon (1919, p. 581), and Tomkeieff (1929, p. 118) have reached the same conclusions in their discussion of gabbroic phases of trap sheets.

A similar relationship between coarse and normal diabase is seen on the east shore of the Connecticut River where it cuts through the Holyoke Range. Veins of coarse pyroxene and labradorite start at the base of the sheet, where there is a torpedo-shaped mass of gabbro about 3 feet long. At the time of the field study the gabbro was exposed down below the river level. A short distance northeast the base of the diabase sheet contains several sedimentary xenoliths which probably continue just below the large gabbro mass.

Most of the gabbroic masses in the Holyoke diabase are flat lenses. It is uncertain whether all of them originated from contact with moist xenoliths. Veinlike masses of coarse diabase are relatively rare in this area, and it is assumed, therefore, that most of the gabbroic lenses somehow have been formed by contact with moist xenoliths, and that these masses, once formed, have been moved along by the simultaneous flow of the lava. Thus they could have acquired their flat, pancakelike shape and become surrounded by normal diabase.

Bain (1941, p. 268) assumed that the Holyoke diabase sheet was extruded from an area at Bare Mountain (west of the Notch) and that part of it flowed eastward, toward the vicinity of Mount Norwottock. This is said to be supported by the curvature of steam holes near

the base of the diabase sheet. Systematic study of this feature has failed to confirm Bain's observation. Figure 6*a* shows the orientation of steam holes measured at 40 localities along the entire floor zone. Spherical or irregular vesicles, within 4–5 feet from the base of the flow, are the most common, but they are by no means ubiquitous. Tubular holes are rare, and most tubes are very nearly perpendicular to the nearest floor—*i.e.*, were vertical before the lava sheet and its floor were tilted. The writer has not been able to discover a flow direction or source area, but it may have been near the eastern border fault. In the vicinity of Millers Falls north of this quadrangle, north-south striking basalt dikes are exposed close to the border fault.

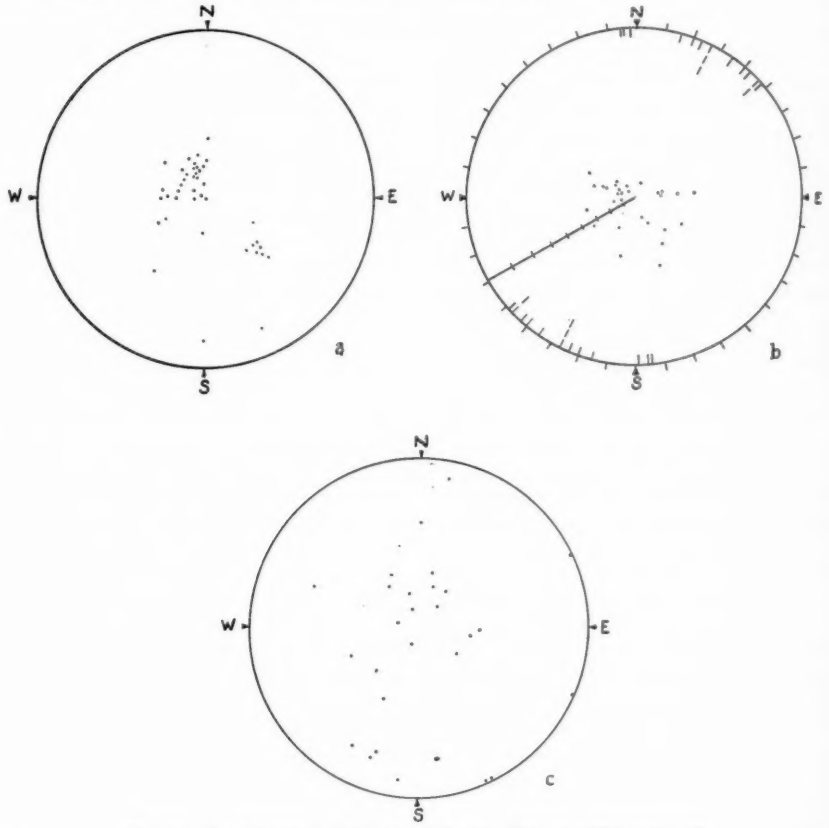


FIGURE 6.—STEREOGRAPHIC PROJECTIONS (LOWER HEMISPHERE)

a. Orientation of tubular steam holes at 40 localities in the Holyoke diabase. The great majority plunge perpendicular to the floor of the sheet.

b. Direction and angle of dip of 46 faults in Mount Holyoke quadrangle. For key, *see* caption of Figure 2. Strike of 15 vertical faults shown by dashes inside the circle. Nearly all faults strike N.–NE., and dip steeply.

c. Bearing and plunge angle of striations on 27 faults.

The famous zeolite localities of New Jersey and Pennsylvania are striking examples of hydrothermal alteration of trap sheets. The alteration of the Connecticut-Massachusetts diabase is less striking and is limited to a few strongly fractured and perhaps faulted zones that cross the diabase sheets in approximately a north-south direction. In this quadrangle the Notch quarry displays the only shatter zone of this kind. The diabase is crossed by irregular, but generally north-trending white veins of

quartz, calcite, and some albite, with minor pockets containing prehnite, datolite, amethystine quartz, a few zeolites, a little fluorite, and, in one place, babingtonite (Shaub, 1941, p. 125; Kitson, 1936).

In the vicinity of the larger quartz-calcite veins the diabase is penetrated by multitudes of thinner, gray, fine-grained veinlets and crack fillings; where they widen, small pockets of quartz, calcite, and sodic plagioclase appear. The diabase between the finer veinlets is brecciated in places, and for several inches in the vicinity of the gray cracks the rock may be discolored to a matted, dull greenish gray. Chlorite-veneered slickensides are also prominently developed in this shatter zone.

Alteration varies in intensity. Most larger outcrops of diabase are crossed by sparse, thin, irregular veinlets that weather out as miniature ridges. They are composed of a fine-grained aggregate of chalcedonic quartz, minor amounts of chlorite, iron oxides, a little albite, and calcite. Veinlets of this kind, more than half an inch thick, are commonly surrounded by brecciated diabase, and the fragments of the breccias have been characteristically altered. Labradorites have been corroded and surrounded by patches of lamellar or plumose albite (An_{2-5}) with variable amounts of quartz. Pyroxenes are invariably chloritized, and patches of magnetite dust have been deposited. Calcite of several generations cuts the chloritized rock, and along the walls of tiny vugs finely fibrous chalcedony forms concentric shells. Table 1 (Nos. 9, 10) gives analyses of two altered rocks. Sodium has increased appreciably, because of secondary albite, and calcite replaces some of the feldspar, but this replacement varies in intensity.

In places, the alteration of diabase has led to a peculiar aphanitic, light-gray rock, either in flat lenses from a few inches to more than 10 feet long, or in spherical masses that lie singly or in groups in greenish-gray, thoroughly chloritized diabase. Emerson (1898a, p. 456) called these rocks limestone inclusions. Indeed, fresh surfaces of the light-gray rock are indistinguishable from lithographic limestone. Touched with cold dilute HCl, they effervesce briskly. But the reaction subsides within seconds, and weathered surfaces of the rock do not effervesce. The rock is largely a mixture of quartz and calcite, with traces of chlorite, sericite, and magnetite. According to an analysis (Table 1, No. 11), part of the carbonate may be dolomite, as there is a surplus of CO_2 over the available lime. A sample was digested in 50 per cent HCl, and the loss in weight showed that calcite amounts to 41.82 per cent of the volume of the rock. A large specimen was broken in two, and one part was digested in HCl for 18 months, until all the carbonate appeared to have been leached. Despite a loss in weight of 39 per cent, the leached sample retained its original shape in minute detail and fits exactly into the irregular surface of separation from the unleached part. A piece of this rock was sent to the late C. N. Fenner, and he stated that during his studies of the Watchung basalts in New Jersey he had seen quite similar "limestone" masses as the final product of hydrothermal alteration of diabase.

Some steam holes near the base of the Holyoke diabase sheet are coated with chalcedony shells, and a central cavity remains. Commonly the innermost layer is crystalline quartz, and no other minerals were noted in the cavities. The agates have formed under conditions different from those that resulted in "limestone." Nowhere have both rocks been seen together, and agate nodules are found in fresh diabase, or alteration is restricted to a narrow shell just outside the agate. Agatized steam holes at the base of the Holyoke diabase are rarely larger than 2 inches, but larger agates weather out at the bottom of a diabase sheet in the Granby tuff, especially south and southwest of Mount Norwottock. These agates have an outer coating of quartz and chlorite, about 3 mm thick, and some of the inner quartz crystals are amethystine.

As the Holyoke diabase sheet is apparently a surface flow, at least within this quadrangle, the source for the altering solutions must be in the volatile fraction of the lava. The albitization of the labradorite of the diabase, as well as the increase of sodium in the two analyzed specimens of altered diabase (Table 1, Nos. 9, 10; Table 2), show that the fluids were relatively sodic. The abundance of chalcedony and quartz in nearly all zones of alteration and breccias shows that much silica was also available. Thus, of the two possible end-stage residua, the sodic-siliceous ones dominated in this area. Never-

theless, iron-rich residua have also been seen in places where magnetite crystals have grown on fracture walls. Good examples are along the northwest contact of the so-called Black Rock diabase breccia and in several places above the floor of the Holyoke diabase sheet. Vesicles in the diabase are lacking within about half an inch of crystal-coated joint surfaces, which is in agreement with views expressed by Fuller (1938; 1950) and Mackin (1947; 1954) that late iron-rich volatiles may escape from cooling magmas or lavas, and that the period of separation partly coincides with the solidification of the crystal mush, so that in the lava vesicles are absent near early-formed fractures. Although, as the association of gabbroic diabase with sedimentary xenoliths shows, some water has probably been imparted to the lava by moisture in the floor, its volume was small compared with that released from the crystallizing diabase. At least there is no suggestion that the floor portions of the sheet are more altered than the top.

As most of the silicified and brecciated zones in the Holyoke diabase trend north-south and dip steeply, it can be assumed that the floor was ruptured and slightly dislocated during the diabase extrusion by incipient faulting, and that the congealing flow was mechanically disturbed by these displacements. There is evidence for at least slight faulting during Longmeadow sedimentation. Rest fluids may have escaped from the cooling lava in fairly large volume where the sheet was transected by zones of crushing and displacement, whereas in less intensely fractured zones the escape was slow and permitted reactions among primary minerals and the sodic-siliceous fluids—a condition similar to that pictured by Shannon (1924) for the cooling of the Goose Creek diabase intrusion in Virginia.

Granby Tuff

The top of the Holyoke diabase sheet is overlain by 12 feet of siltstone (exposed at Dry Brook and also in the extreme southwest corner of the quadrangle), which is succeeded by sandstone and conglomerate totaling about 140 feet. In the western part of the quadrangle, this basal portion of the so-called Longmeadow sandstone is 140 feet thick and is overlain by a series of tuffaceous rocks, named by Emerson (1898a, p. 369) the Granby tuff.

Outcrops of Granby tuff consist of chocolate-brown, fine-, medium-, or coarse-grained, well-stratified, or massive layers. In many places tuffaceous lenses pass laterally into silty sandstone, into basaltic pyroclastics, and even into massive diabase. The tuff has its greatest thickness (about 500 feet) near Smiths Ferry, on the west side of the Connecticut River, and it thins eastward by wedging and lensing so that at the eastern quadrangle border there is no more than 10–12 feet of tuff. As the tuff series thins, the underlying Longmeadow sandstone and conglomerate thickens to about 500 feet, and perhaps to as much as 1200 feet. Exact figures are difficult to state because of the lenticular and cross-bedded nature of the sandstone.

Under the microscope the tuff reveals the same gradation from igneous to sedimentary types. Igneous phases are essentially like Holyoke diabase, but the amount of chloritized pyroxene exceeds that in fresh Holyoke diabase. It may be as much as 15 per cent of the mafic minerals in otherwise nearly fresh rock. Indices of refraction and extinction angles of both plagioclase and pyroxene, determined in several specimens, fall within the limits of the Holyoke diabase. In the better-stratified tuff, the minerals are more extensively altered: pyroxenes are replaced by chlorite, iron oxide masses, and calcite; feldspars turn to cloudy aggregates of sericitic mica, clay minerals, and carbonate. Where the material has been transported farther and mixed with surface debris, more quartz and potassium feldspar enter the diabasic mineral suite, accompanied by minor amounts of zircon, tourmaline, garnet, muscovite, bleached biotite, and clastic magnetite. These mixed types of tuff are lighter shades of brown, purple, and maroon. Coarse pyroclastic rocks, with basaltic bombs, lenses of massive diabase, diabase breccia, intermixed diabasic ashes, and blocks of sandstone, are exposed along the Holyoke-Northampton highway on the west bank of the Connecticut River, southeast of Kennedy Pond, and in cuts of the same highway directly south of the southern quadrangle border. Other outcrops of coarse pyroclastic

rocks are north-northeast of Aldrich Lake, hill 619′ (1 mile south-southeast of Mount Norwottock), and northeast of Moody Corner.

Sedimentary phases of the Granby tuff, although in composition like other sandstone and conglomerate layers of the Longmeadow formation, retain a darker, brownish color that is due to dark alteration products of pyroxene. Microscopically all rocks show thin films or patches of decomposed diabase granules in which remnants of lath-shaped plagioclase are faintly visible surrounded by mushy aggregates of clay minerals, stained dark by limonite, and chloritic masses derived from pyroxene and magnetite.

A diabase flow 18–30 feet thick, the Hampden diabase of Emerson (1898a, p. 464), is prominent near the base of the tuff sequence southeast of Mount Holyoke. One mile north-northwest of Pearl City Pond it overlies thin-bedded tuff, but at the north-northwest end of hill 531′ it rests on sandstone, which shows that the pre-diabasic tuff was distributed irregularly. At the southeast base of Mount Holyoke, no more than 110 feet of sandstone, conglomerate, and tuff separates the top of the Holyoke diabase and the Granby flow, but south-southeast of the Notch, the flow lenses out and rests on a very thick sandstone-conglomerate mass. Thus, after the extrusion of the Holyoke diabase flow, and before eruption of the Hampden flow, a thick alluvial fan of coarse gravel and sand—the Longmeadow sandstone—had grown from an eastern source, thinning out near the longitude of the present

TABLE 2.—MODE, NORM, AND CATION PERCENTAGES OF FRESH DIABASE (Table 1, nos. 1, 7), AND ALTERED DIABASE (Table 1, no. 9)

Calculation of cation percentages follows Eskola (1954). Lower right-hand table gives cation changes from fresh to altered diabase. Plagioclase in mode includes small amounts of interstitial glass, with albite and potassium feldspar, and traces of quartz.

Analysis 1

mode		norm			cation percent	O in oxides
plag.	48.98	quartz	4.22	SiO_2	49.458	98.516
pyrox.	40.98	orth.	4.45	TiO_2	.812	1.624
magn.	5.48	alb.	21.48	Al_2O_3	14.904	22.356
chlor.	4.62	anorth.	22.24	Fe_2O_3	2.562	3.285
		woll.	10.90	FeO	7.745	7.745
		ferros.	12.94	MnO	.161	.161
		enst.	14.70	MgO	8.442	8.442
		magn.	5.10	CaO	10.164	10.164
		ilmen.	2.13	Na_2O	4.757	2.378
		apat.	.34	K_2O	.904	.452
			98.50	P_2O_5	.097	.242
		H_2O	1.31		100.006	155.363
		CO_2	.30	H_2O^+	(3.933)	- 3.933
			100.11			151.432
		normative plagioclase: $ab_{59}an_{51}$				+ 7.866
						159.298

Analysis 7

mode		norm			cation percent	O in oxides
plag.	40.24	quartz	1.62	SiO_2	48.502	97.004
pyrox.	46.38	orth.	3.89	TiO_2	1.024	2.048
magn.	10.01	alb.	19.91	Al_2O_3	15.13	22.695
chlor.	3.69	anorth.	23.35	Fe_2O_3	2.384	3.576
		woll.	10.21	FeO	9.364	9.364
		ferros.	17.42	MnO	.17	.17
		enst.	14.00	MgO	7.593	7.593
		magn.	4.64	CaO	10.306	10.306
		ilmen.	2.58	Na_2O	4.614	2.307
		apat.	.34	K_2O	.77	.385
			97.96	P_2O_5	.141	.353
		H_2O	1.17	H_2O^+	(3.208)	
		CO_2	1.56			
			100.69		99.998	155.801
		normative plagioclase: $ab_{46}an_{54}$				- 3.208
						152.593
						+ 6.416
						159.009

Analysis 9

norm			cation percent	O in oxides
orth.	3.34	SiO_2	47.560	95.120
alb.	35.11	TiO_2	.993	1.986
anorth.	13.90	Al_2O_3	14.337	21.505
woll.	9.40	Fe_2O_3	2.294	3.441
ferros.	6.20	FeO	9.201	9.201
enst.	4.50	MnO	.145	.145
forst.	7.42	MgO	8.771	8.771
fayal.	11.42	CaO	7.778	7.778
magn.	0.70	Na_2O	8.063	4.031
ilmen.	2.58	K_2O	.737	.368
apat.	.34	P_2O_5	.116	2.901
	94.91		99.995	155.247
H_2O	3.30	H_2O^+	(3.962)	- 3.962
CO_2	.15			151.285
	98.36			+ 7.924
normative plagioclase: $ab_{69}an_{32}$				159.209

more in fresh diabase (analysis 1)		more in altered diabase (analysis 9)	
1.898	Si	.181	Ti
.567	Al	1.456	Fe^{2+}
.268	Fe^{3+}	.329	Mg
.016	Mn	3.306	Na
2.386	Ca	.019	P
.167	K	5.291	
5.302			

Connecticut River. This is evidence that rapid sedimentation and volcanism proceeded contemporaneously in this area.

The upper contact of the Granby diabase flow is sharp against sandstone in some places, but in the other places the flow is overlain by basaltic tuff, in which case the rocks commonly grade into each other, as they do north-northeast of Aldrich Lake. Here the base of the tuff is blocky and massive, but layering becomes visible near the top of the tuff which is about 40–60 feet thick. South of Mount Norwottock, on hill 619′, the tuff contains blocks of vesicular diabase without sharp borders against the tuff.

The Granby tuff and adjacent sections of the Triassic sedimentary rocks are cut by numerous pipes, dikes, and sills of fine-grained diabase. Seventy-eight intrusive bodies have been marked on Plate 1, and several smaller ones were seen. Most of them are pipes that rise as small crags, 10–30 feet above the surrounding tuff surface. Where exposed, the borders are about vertical, and diameters of the pipes range from 15 to 200 feet. Discordant diabase dikes are associated with the Hampden diabase flow southeast of Taylor Notch; small, irregularly shaped crosscutting apophyses occur southeast of Mount Holyoke and north-northwest of Aldrich Mills. Sills, conformable with the bedding in the sandstone and conglomerate below the Granby tuff, are scattered through the hills about 1–1½ miles northeast of Moody Corner. Possibly some dikes and sills are connected with pipes, but no contact exposures offer conclusive evidence.

The pipes are composed of relatively dark, fresh, splintery, locally brecciated diabase, holding angular fragments of slightly lighter-colored diabase (analysis, Table 1, No. 7). Microscopically, the diabasic texture is less pronounced than in the Holyoke diabase, but in both the plagioclase is somewhat earlier than the pyroxene and is seriate from sparse phenocrysts 1–2 mm long to groundmass laths $\frac{1}{6}$–$\frac{1}{100}$ mm long. In specimens from seven pipes the plagioclase is labradorite (An_{54-69}, mostly An_{62}). Pyroxene in the finer-grained pipes is almost invisible, as the minute grains have low interference colors. Some vesicles filled with chlorite are surrounded by radial groups of colorless pyroxene, attesting to the relatively late growth of some of the crystals. A small amount of pale-brown glass was noted in pipes southeast of Lake Bray and in the dike 1 mile northeast of Dry Brook Hill. Olivine, reported from there by Emerson (1898a, p. 493), could not be identified in four specimens and thin sections. A chemical analysis by *semimicro* methods is listed (Table 1, No. 8). At the contact of this dike with Triassic sandstone, a little albite (An_8) has entered the sandstone, and biotite plates, $\frac{1}{20}$ mm across, have grown apparently at the expense of pre-existing chlorite in the matrix of the sandstone. This is the only contact effect noted in the quadrangle. Quartz grains were seen in several thin sections of pipe diabase. They are probably material taken up from the underlying Triassic sediment. Potassium feldspar, however, was not seen; if it is present, it must have been comminuted.

A large northeast-trending dike of diabase breccia extends from the upper part of Dry Brook into the upper portion of the Holyoke Range, south of Taylor Notch. To it Emerson (1898a, p. 489–494) gave the name Black Rock dike, in reference to the local name of a promontory which this dike makes on the back slope of the diabase ridge (Fig. 3). One mile northeast of its northeast end, the smooth till cover on the north slope of the Holyoke Range is interrupted by innumerable small, scattered ledges of diabase breccia.

The cliffs of the Black Rock dike display a complex of vesicular, irregularly cemented diabase fragments, ashes, and foreign fragments, penetrated by lenses of firm diabase showing imperfect columnar jointing in places. Under the microscope the dark, fine-grained, iron-stained groundmass of the breccia is seen to hold many crystals of altered labradorite, but in places albite and oligoclase were also noted. Pyroxenes are largely replaced by secondary iron oxide and chlorite, and patches of quartz and calcite penetrate the entire mass. Clastic admixtures of quartz, potassium feldspar, microperthite, zircon, and muscovite are widespread. Depending on the degree of alteration and foreign admixtures, a diabasic texture remains discernible, and, even where most of the primary silicates have been altered to clay and carbonate, a faint fluxion structure of the outlines of plagioclase laths survives.

In contrast to the pipes, the Black Rock

breccia contains many fragments of Triassic sediments, especially in the northeastern ledges. They are easily identified with the Sugarloaf arkose, and, as pointed out by Bain (1941, p. 267), the arkose was apparently unconsolidated at the time of the eruption, for single flesh-colored potassium feldspar fragments are embedded in about all diabase outcrops, and the outlines of larger fragments are rounded. In an outcrop of diabase breccia directly north of the north base of the Holyoke diabase sheet, the firm breccia diabase is "cut" by a vertical dike of pink arkose, 4 inches thick and at least 5 feet high. This grit must have been squeezed as a soft paste into a newly formed fracture in the cooling breccia.

The Black Rock diabase cuts the basal sediments above the Holyoke diabase sheet, and also basal members of the Granby tuff, but its relation to the diabase flow in the Granby tuff is in doubt. The Black Rock breccia is probably a particularly large dike that was emplaced at about the same time as the pipes and dikes that cut the tuff. The breccia follows closely a zone within which pipes and dikes on the west side of the Connecticut River were emplaced. The Holyoke diabase sheet resisted the pressure of the breccia, but in the incoherent Sugarloaf arkose a system of interconnected channels and cracks developed in which the lava mingled with arkosic debris over a large area. The Black Rock breccia may mark the location of an important fissure feeder for the late diabase. A second, less productive zone branched off eastward from the Black Rock fissure. The same zone could be responsible for the eruption of the Holyoke diabase sheet, but evidence is inconclusive.

Longmeadow Sandstone

The youngest subdivision of the Triassic System in this quadrangle, the Longmeadow sandstone, was named by Emerson (1898a, p. 364–369) in reference to quarries in Longmeadow township, near Springfield, where the rock was quarried as a building stone, the popular "brownstone" of the last century. As the Mount Holyoke quadrangle is northwest of the center of the southern Triassic basin, only the lower and middle parts of the sandstone formation crop out within its confines, and the latest Triassic sediment, the Chicopee shale, is not known to be preserved within Mount Holyoke quadrangle.

The best outcrops of the Longmeadow sandstone are south of the Holyoke Range, east of the paved road from South Hadley to Amherst; other good outcrops are west of that road, about 1 mile southwest of the Notch; in the Mount Tom State Reservation; and on the east bank of the Connecticut River, southeast of Smiths Ferry.

As already mentioned, the earliest beds of this formation are older than the Granby tuff, but the middle and upper parts of the Longmeadow sandstone rest above the tuff. The Longmeadow sandstone and conglomerate are much thicker on the east side of the Triassic basin and tend to wedge out westward, as indicated by the distribution of coarse and fine debris in the formation (Pl. 2) and by the prevailing southwesterly dip of fore-set beds in cross-bedded phases of the formation (Fig. 2). The chief rock types that occur as pebbles are various granites, pegmatite, vein quartz and some rarer fine-grained quartz schists and quartzose amphibolites. Phyllite and slate pebbles would be more abundant had the rocks west of the Connecticut River furnished much of this debris.

The sandstone varies from pale-yellowish gray to a rich maroon or russet. Many of the coarser, conglomeratic lenses are yellowish gray whereas finer-grained micaceous sandstone beds are reddish. Because of the lenticular nature of the strata, cross-bedding, and flattening of the beds toward the interior of the basin, the true thickness of the Longmeadow sandstone in this area is indefinite, but 2000 feet is estimated.

The Longmeadow sandstone has furnished many dinosaur footprints, as well as scattered bones (Lull, 1912; 1915; 1917) and an almost complete skeleton of a small dinosaur, *Podokesaurus holyokensis* (Talbot, 1911). A short distance north of Aldrich Mills (Pl. 1) innumerable footprints are crowded on a sandstone bed. A short distance south of the quadrangle border, on the west bank of the Connecticut River, is a large bedding-plane surface of the sandstone with hundreds of excellently preserved footprints of various species of dinosaurs. Fortunately, the ledge is now legally protected

as a State Monument. A third sandstone ledge with dinosaur footprints is on the property of Miss Helen Griffith, in the northwest part of South Hadley. Poorly preserved stem fragments of *Equisetum*, ripple marks, mud cracks, and rain-drop impressions are seen in some outcrops of the sandstone.

Fine-grained sandstone beds consist of small quartz grains with subordinate feldspar, and muscovite flakes cover the principal bedding planes. Orthoclase, microperthite, and a sodic oligoclase are the principal feldspars, and among the heavy minerals zircon, tourmaline, magnetite, garnet, sphene, biotite, and apatite have been identified. Partly chloritized pyroxene and, very rarely, basalt fragments have also been noted. The cement is for the most part carbonate, with subordinate iron oxide, silica, and clay. Very locally, small blebs of chalcedonic quartz and chert are seen.

Although the Chicopee shale is lacking in the quadrangle, some thin lenses of dark-brown, almost black siltstone and sandy shale are interbedded with fine-grained sandstone near the southern border and presumably represent scattered feather edges of the lake in which the Chicopee shale was deposited. One outcrop is south-southwest of Granby, and in 1946, during excavations for the heating plant of Mount Holyoke College, slabs of purplish-black shale were exposed. A fish fauna has been described from localities in both the northern and southern Triassic basins, and in some localities just south of the quadrangle (Eastman, 1911).

Quaternary Deposits

General Statement

As elsewhere in northeastern United States, the interval between the Triassic and the Quaternary periods was a period of prolonged erosion, during which the present landforms were essentially established. During the Pleistocene epoch the area was glaciated, and the deposits resulting from glaciation and subsequent melting stages of the ice are well represented. It is unnecessary to review here the literature on this subject, for the mapping of a relatively small area, even though fairly detailed, could not solve large problems of regional Pleistocene geology. For a background of the Pleistocene and post-Pleistocene history of the region, the papers by Antevs (1922; 1925), Jahns and Willard (1942), Flint (1930; 1933), and Lougee (1953) are helpful.

The following units of Pleistocene and post-Pleistocene ages are recognized in the area: (1) glacial till, (2) outwash deposits formed under and around melt-water lakes whose areas and levels changed from time to time, (3) post-Pleistocene alluvium of several kinds.

It is in the nature of these unconsolidated deposits that few of their contacts are well exposed. The reconstruction of the late glacial and postglacial history is, therefore, to some extent inferential. In the Connecticut Valley the sequence of events worked out by Jahns and Willard (1942) is about as reliable an account as can be reconstructed at present.

Glacial Till

Deposits of only one till sheet, probably the Wisconsin, could be identified in the quadrangle. It forms a thin but persistent cover over the crystalline bedrock in the northeastern part of the quadrangle, forms many low hills in the southern area, but is nearly absent from the Holyoke-Mount Tom range. Though accurate measurements were impossible, its thickness is thought not to exceed 30 feet. Thicker till cover may, however, overlie preglacial depressions.

The till is poorly exposed; shallow cuts in a few ravines on the north slope of the Holyoke Range and scattered showings in the lowland southwest of Amherst indicate that it is fairly clayey and compact and is relatively resistant to present-day erosion. No stratification was seen. The bedrock floor beneath the till is grooved in places, and Plate 2 shows the direction of the striae. As elsewhere in this part of New England (Flint, 1945), the general trend is north-northwest, but, where the ice passed the narrows between Mount Nonotuck and Mount Holyoke, the trend shifted to northwest, and the ice seems to have flared outward in diverging directions, once it had passed the gorge.

The ice sheet moved across the steep front of the Holyoke Range without deflection, as shown by the striae on the range which retain a uniform south-southeast direction. Nothing suggests that the ice in contact with the diabase

wall moved parallel to the wall. Thus we have independent evidence that the ice sheet was thick enough to surmount the 700-foot wall of the Holyoke Range without shifting its direction. The striae elsewhere in New England and the Adirondacks also display uniform direction, without regard to present relief.

Eskers

Typical eskers are lacking in this quadrangle. A short, curving gravel ridge, 1 mile due east of Moody Corner and south of Aldrich Lake, resembles one. This could be a crevasse filling, or the fill of a short englacial stream channel. The many kettle holes in the surrounding gravel and sand plain attest the gradual dissipation of ice masses in the vicinity of the esker.

Late Pleistocene Outwash and Lake Deposits

Overlying the till, generally at low elevations, are various gravel and sand deposits, the outwash from meltwater streams or sheets that escaped from the hill areas toward the valleys during the end of the Glacial stage. The succession of deposits along the north slope of the Holyoke Range is fairly typical. A discontinuous, narrow, moundlike deposit of coarse, ill-sorted kame gravel that fringes the till-covered slope merges southward into till and northward into better-sorted, finer gravels. Whether the coarse kame gravel was originally flat-topped cannot be decided, as this deposit is being actively dissected now by short ravines that drain the north slope of the range.

At about 200–180 feet the gravels are finer, and their surface flattens out until, at elevations between 180 and 160 feet, a sand terrace rests as a wedge, 0 to 30 feet thick, upon varved clays, the typical lake-bottom deposits of the valley. Antevs (1922; 1925; 1930–1932) gives a full description of these deposits. An excellent exposure of the clays is at the east bank of the Connecticut River, half a mile south of Hadley (Jahns and Willard, 1942, Pl. 3, B).

The top elevation of the varved clays could be determined at 47 localities and was mapped throughout the quadrangle. The figures of Plate 2 show that the surface slopes gently from the uplands toward the axis of the Connecticut Valley. For example, from the bluff east of the mouth of Fort River, at about 160 feet, the top of the clay rises eastward to about 195 feet (1 mile south-southeast of Harts Brook School). Similarly, along the east-west course of Fort River, the top of the clay rises from about 140 feet on the west to over 210 feet (1 mile north of Devils Garden). Corresponding rises, away from the axis of the Connecticut Valley, are seen elsewhere in the area (Pl. 2).

Jahns and Willard (1942, p. 286) describe localities where the gradation of varved clay into silt and sand of the overlying wedge of coarser sediment can be seen. When the field work was done, a good contact was exposed on the east side of the Connecticut River, southeast of Shepherd Island, where a power-transmission line crosses a local ravine. Farther north in the valley, the writer has seen other contacts in which the interfingering of varved clay, silty sand, and clayey silt is well displayed. All the facts known to the writer support Jahns' and Willard's conclusion that the gravel, sand, and clay deposits are contemporaneous, and that the relatively coarse debris forms wedge-like aprons that connect the shore zones of the former meltwater lakes with the deeper-bottom zones where only varved clay was deposited.

The differentiation of meltwater lake deposits into gravels, sands, silts, and clays, which went on in the main valley on a large scale, had its miniature replica in several detached, or partly detached, smaller basins. Thus, the narrow depression between South Amherst and Mount Pollux, mainly underlain by kame gravel and sand, has in its central, lowest area typical varved clay, which was exposed during the writer's field work. This deposit has an upper surface 40–70 feet higher than that of the two larger lobes of clay that were connected directly with the Connecticut Valley. Similar small detached basins existed east of South Hadley and east of Thermopylae, at the south base of the Holyoke Range.

How much sand and gravel was contributed in each locality by local streams and rills of meltwater, and how much was added by transcurrent meltwater streams is impossible to determine. The gravel and coarser sand came perhaps largely from local till areas, but some of the finer silt and clay may have had a distant source. Each local area has certain peculiarities

that can be understood best if the immediately surrounding topographic features, the nearest channels of water, local relief, and the presumably irregular distribution of ice remnants are taken into consideration. The incipient erosion of the deposits in Recent time has further added to the difficulties of reconstructing the details.

In general, the superposition and orderliness in mutual relations among the various types of deposit are most uniform where the present surface slopes gently and uniformly toward the axis of the Connecticut Valley. An example of a symmetric deltaic deposit, resting on varved clay, is between South Hadley and the Connecticut River. The apex of the delta is at 230 feet elevation. The top of the clay rises toward it from 160 feet, near the river, to 180 feet, 1 mile from it, so that the deltal sand thickens from zero feet, nearest the Connecticut River, to about 50 feet at the apex of the delta.

In places rather coarse gravel masses border the Connecticut River yet are found at elevations of 290 and 270 feet, and are believed to have formed at an early stage of the meltwater period. The largest, 2–3 miles west-southwest of Amherst, forms a broad barrier extending from Mount Warner toward the bedrock area at Amherst. The second, Dry Brook Hill, southeast of Thermopylae, suggests deposition by a large flow of water that had broken a barrier. Up to that stage the "Narrows," where the Connecticut River crosses the Holyoke Range, was filled with till and ice. When it gave way, the excess debris should have been dropped below the breach as soon as the velocity of the torrent subsided. Dry Brook Hill is about half a mile below the Narrows, and the beds in all exposed parts of the gravel dip 10°–28° southeastward; the current that deposited the gravel flowed southeastward. The gravel of Dry Brook Hill is older than the clay and terrace sand deposited along its eastern flank. At the northwest base of the hill a mass of purplish clay appears in the shore of the Connecticut River, at an elevation of about 105 feet, dipping at about 15° SE. into Dry Brook Hill, and showing intense twisting and squeezing. This could be a remnant of an early clay deposit that was contorted and plowed into by the mixture of ice and debris when the dam broke.

A third mass of early gravels, just south of Harts Brook School, occurs in a sand-covered clay area; its mode of deposition and source area are obscure. The red and pink color of these three gravel masses shows that they have been derived mainly from local till deposits, for till that is derived from crystalline rocks is grayish.

Additional masses of relatively coarse and, presumably, old gravels are poorly exposed in the lowland south of Amherst. They have been called kame gravels, mainly because they lack surface expression or significant form, are adjacent to till, and lack sharp borders against till. However, better contact exposures might change this interpretation. Many details in the distribution, composition, and shape of sand and gravel deposits have been studied by R. R. Wheeler (1941).

The precise conditions that led to the draining and extinction of the meltwater lake of the Hadley-Holyoke area cannot be ascertained through the study of a small local area. According to Jahns and Willard, the channel near New Britain, Connecticut, which first interested Loughlin (1905, p. 24–25), may have been the point where a till barrier that had sustained the meltwater lake level, maintaining one, or several interconnected lakes, finally gave way. With the draining of this lake the late Pleistocene period of deposition ended. The newly exposed lake floor was dotted by numerous kettle holes where detached or grounded masses of ice were wasting away while clay, silt, and sand were being deposited around them. Examples are east and north of Aldrich Lake.

Recent Deposits

Deposits formed since the draining of the late-glacial lake consist of insignificant patches of silt and sand along the courses of streams that have subsequently incised shallow gullies and valleys into the Pleistocene deposits. Large surfaces underlain by alluvium are, however, found in the flood plain of the Connecticut River. As the distribution of these deposits is of the greatest importance for planning settlements, highways, and railroads, for instance, a special study of these deposits was made, after the disastrous flood of March 1936 by R. H. Jahns (1947). He stated (1947, p. 59) that

during the 1936 flood four-fifths or more of the entire postglacial flood plain was covered by water. From north to south, the following levels of the flood crest were determined in the quadrangle:

	(Feet)
Hatfield	130
Hadley	129
Coolidge Bridge, Northampton	128
Highway Northampton-Holyoke, half a mile south of Northampton	128
Half a mile above Smiths Ferry	118

The flood deposits are composed commonly of a few inches of silt, though the thickness may increase in places, for instance near scour pits where eddies churned up the water and caused irregular patterns of deposition (Collins and Schalk, 1937). Locally, much coarse debris, driftwood, and vegetation were mixed with the flood deposits.

The prevailing westerly winds redistribute the finest detritus of the Connecticut Valley flood plain in small, shifting dunes, of which a number have been indicated on the east side of the river (Pl. 1).

In several swamps and bogs, minor amounts of black muck and impure peat, mixed with silt, are being deposited.

Structural Geology

General Statement

As the Quaternary deposits cover most of the quadrangle, the structure of the consolidated bedrock formations is imperfectly known. The few exposures of crystalline rocks suggest a west- or northwest-dipping series of schists, presumably the continuation of the west slope of the large system of domes that extend, with north-trending axes, along the east flank of the Connecticut Valley of New Hampshire, Massachusetts, and Connecticut (Willard, 1952b).

The irregular floor beneath the Triassic sediments has been noted, as well as the basin structure of the Triassic rocks in the southern part of the quadrangle. Plate 2 shows the general synclinal structure of the Triassic rocks. It remains to add a few details concerning faults that are known or suspected to cross the Triassic rocks.

Faults

Plate 1 shows about 12 faults, either exposed, or necessary on account of known or visible displacements of identifiable horizons. Almost all of them strike north-northeast, dip steeply or are vertical, and the displacements along them must be measured in terms of a few feet rather than in hundreds of feet. Figure 6*b* gives the orientation of 46 measured fault planes, and Figure 6*c* the orientation of striations on 27 of them. Nearly all the striae plunge at fairly high angles, either to the northeast or southwest. On a few faults striae in more than one direction appear, adding to the difficulty of interpreting correctly the net result, or net slip. It is unfortunate that the apparent horizontal separations shown on Plate 1 occur on faults that fail to show striations as Bain (1941, p. 269–272) has advocated considerable horizontal slip in the area east of Mount Holyoke quadrangle. His description is so sketchy that it is impossible to verify the data. Some of the faults in Mount Holyoke quadrangle may have had mainly horizontal displacements, others perhaps steep to vertical ones. It is probable, for instance, that the diabase block east of the Notch, with a summit about 200 feet lower than its neighbor peaks along the crest to the west, has been relatively depressed by faults. Internal structure and thickness of the diabase sheet are not necessarily uniform from one end of the range to the other, however, and it could be that the Notch area was farther from the nearest center, or centers, of eruption than the adjoining hills and so received from the beginning a thinner lava cover.

There is convincing evidence that a broad fault zone of north-northeasterly strike dissects the Triassic terrain just west of the village of Granby. As Plate 2 shows, the strikes in the Longmeadow sandstone vary abruptly here. West of Porter Swamp, they strike west-northwest; east of it, they strike north-northeast. The fault zone itself is concealed by Quaternary deposits. It obviously represents the junction of two hinge faults or, rather, a fracture zone along which two blocks join that have rotated in slightly different directions. The eastern block has rotated down to the southeast, the western one down to the south-southwest. Both movements can be explained as accom-

modating the gentle sagging of the southern Triassic basin. As this is supported by a crystalline floor of heterogeneous composition and, presumably, structure, the two blocks may be presumed to ride on two adjacent crystalline blocks that were tilted in slightly diverging directions, though each has moved so that the center of the basin was relatively depressed by the rotation. But the centers of depression are not quite the same for each block.

The motion of these two blocks appears to have died out toward the Holyoke Range. However, the easternmost block, southeast of Mount Norwottock, still exhibits the northeast strike of the sediments in contrast to the block next west, in which the beds trend east-west. The mapped fault along the north-northeast trending depression due south of Mount Norwottock is probably one branch of the Granby fault zone.

Bain (1941, p. 265) is of the opinion that faulting of Triassic age aided the development of the basin structure. In a ledge of conglomerate on a small hill, 370-foot elevation, 1 mile north-northeast of Moody Corner, some of the beds are faulted as much as a foot, whereas the top beds in the ledge are unfaulted.

There is no evidence in this quadrangle of any crustal movements in post-Triassic time. However, in a few localities in the Connecticut Valley varved clays are slightly tilted to the south, and independent evidence suggests strongly that southern New England has been slightly tilted downward to the south since the recession of the ice sheet.

Ground Water

The survey investigated the distribution of ground water. At present the area has no pressing water-supply problems, but demand may well increase in the future.

Most of the towns and villages obtain ample water from reservoirs, several of which are on the south and north slopes of Holyoke Range and Mount Tom Range. Most farms removed from villages and dairies have wells of their own, and the quality of water is satisfactory. The most important shallow-water horizon is the top of the varved clays. As it underlies most of the Connecticut Valley, wells drilled through no more than 15–40 feet of sand may strike this horizon. The approximate depths to this aquifer are outlined by dashed contours on Plate 2. Few wells have struck adequate amounts of water in the impervious till. The Atkins orchards near the top of Mount Pollux obtain ample water from a well 120 feet deep, which presumably has struck a kame gravel pocket in or above the till; yet two wells N. James Schoonmaker put down at the west base of Mount Pollux to 69 and 27 feet, respectively, and bottomed in bedrock, are dry.

The Triassic basins are large and as yet almost untapped reservoirs. Should the needs for ground water increase greatly, deep wells are likely to yield large flows from these rocks. Several paper factories in the city of Holyoke have resorted to such wells for supplementary supplies. Water in the deeper parts of the sandstone basins should stand under artesian pressure, and some of the aquifers seem to be replete with ground water, as is seen at the north slope of the Holyoke Range. Between the Notch and Mount Holyoke, there are at least four water seeps from Sugarloaf arkose, directly below the diabase sheet. The arkose bed is probably so filled with water that the excess spills out over the edge of the basin. Particularly large flows may be available where faults or shatter zones have loosened the rocks along the rim of the basin; the large spring that supplies the crusher of the Notch quarry with water may be of this kind. It is located where one of the north-south zones of brecciation in the diabase intersects the floor of the sheet.

A large spring half a mile north-northwest of Granby issues at the base of a cuesta of east dipping conglomerate and is probably located on the east side of the Granby fault zone. The water appears to be under artesian pressure.

Soils

The careful and detailed soil survey of Hampden and Hampshire counties, Massachusetts, by Latimer and Smith (1929) was unknown to the writer until after the field work had been completed. The formation boundaries on the geological map agree closely with the soil-type distribution, as mapped by these authors. It is hoped that the mapped distribution of the

various bedrock and Quaternary formations will, in turn, prove useful to further investigations of soil types in this part of New England.

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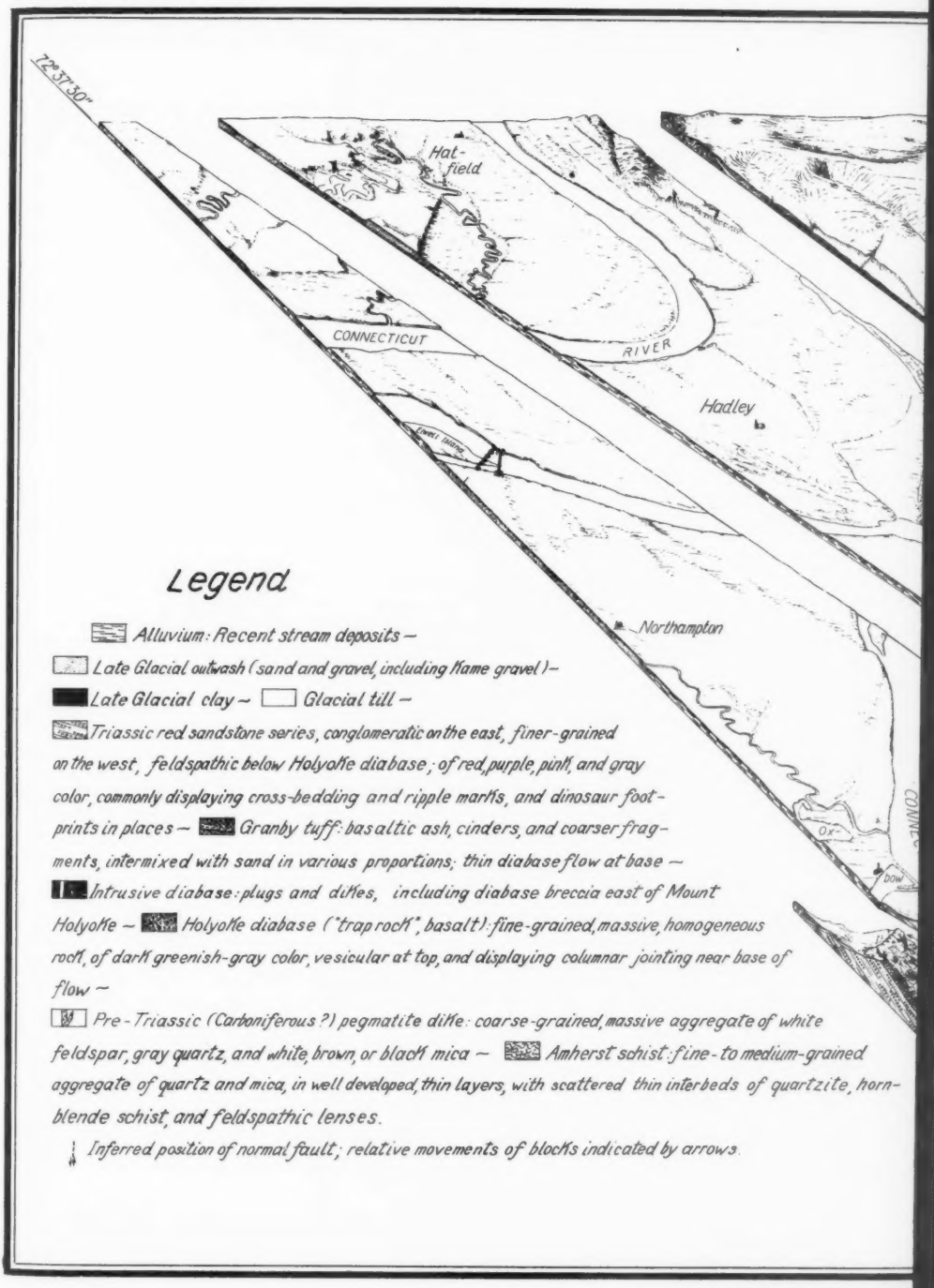
Manuscript Received by the Secretary of the Society, September 3, 1954

Project Grant 618/53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Hat-
field
CONNECTICUT
RIVER
Hadley
Northampton
Legend
Alluvium: Recent stream deposits —
Late Glacial outwash (sand and gravel, including Kame gravel) —
Late Glacial clay — Glacial till —
Triassic red sandstone series, conglomeratic on the east, finer-grained on the west, feldspathic below Holyoke diabase; of red, purple, pink, and gray color, commonly displaying cross-bedding and ripple marks, and dinosaur foot-prints in places — Granby tuff: basaltic ash, cinders, and coarser fragments, intermixed with sand in various proportions; thin diabase flow at base —
Intrusive diabase: plugs and dikes, including diabase breccia east of Mount Holyoke — Holyoke diabase ("trap rock", basalt): fine-grained, massive, homogeneous rock, of dark greenish-gray color, vesicular at top, and displaying columnar jointing near base of flow —
Pre-Triassic (Carboniferous?) pegmatite dike: coarse-grained, massive aggregate of white feldspar, gray quartz, and white, brown, or black mica — Amherst schist: fine- to medium-grained aggregate of quartz and mica, in well developed, thin layers, with scattered thin interbeds of quartzite, hornblende schist, and feldspathic lenses.
Inferred position of normal fault; relative movements of blocks indicated by arrows.

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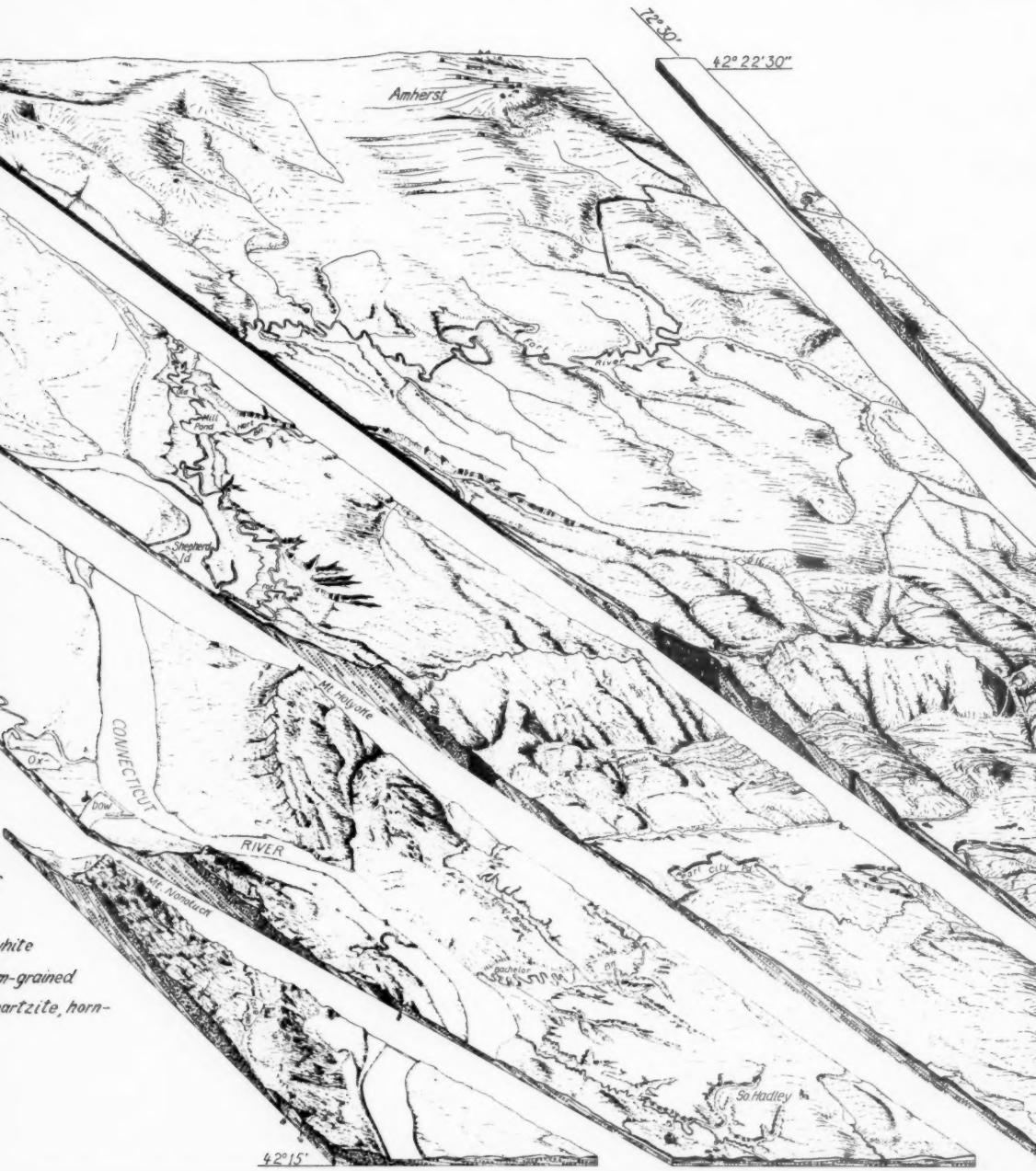

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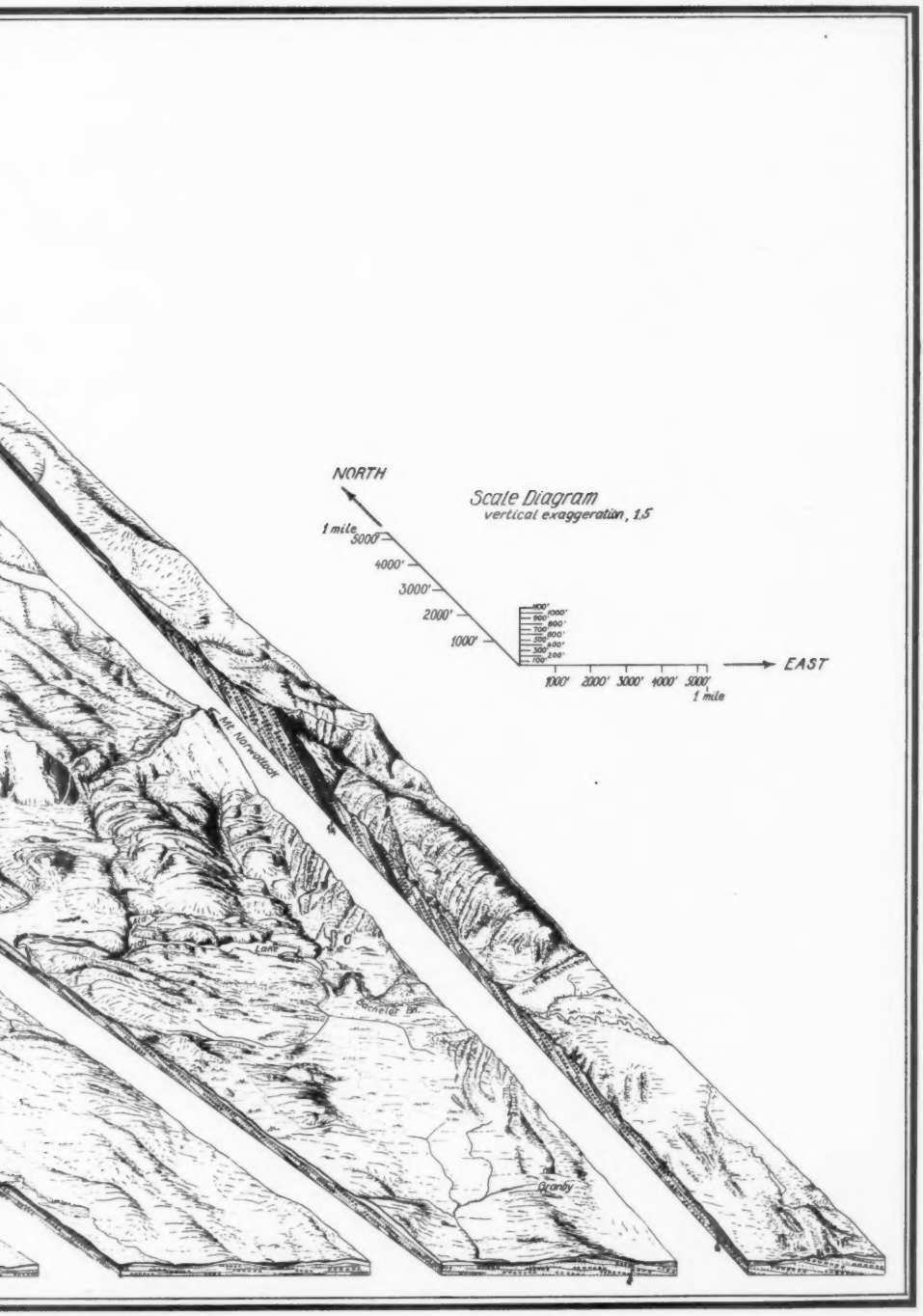
OGIC RELIEF MAP OF MOUNT HOLYOKE QUADRANGLE, MASSACHUS


NORTH
Scale Diagram
vertical exaggeration, 1.5
1 mile
5000'
4000'
3000'
2000'
1000'
EAST
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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 68, PP. 505–526, 10 FIGS. APRIL 1957

LITHOFACIES OF THE SALT WASH MEMBER OF THE MORRISON FORMATION, COLORADO PLATEAU

By THOMAS E. MULLENS AND VAL L. FREEMAN

ABSTRACT

The Salt Wash is the basal member of the Upper Jurassic Morrison Formation in parts of Utah, Colorado, Arizona, and New Mexico. Deposited by streams, it comprises lenticular beds of cross-laminated sandstone irregularly interbedded with mudstone, siltstone, claystone, and horizontally laminated sandstone.

The term "lithofacies," as used in this paper, denotes lithologic aspect. The specific lithofacies of the Salt Wash member at a given locality is determined by the thickness, proportion, and continuity of the stream and flood-plain deposits that make up the Salt Wash. Stream deposits include all rocks interpreted as deposited from moving water; flood-plain deposits include all rocks interpreted as deposited from slack water.

Regional differences in lithofacies show that the Salt Wash member is a fan-shaped wedge of sedimentary rocks whose apex is in south-central Utah. Within the wedge, the thickness of the Salt Wash and the thickness, proportion, and continuity of the contained stream deposits decrease relatively uniformly to the north, northeast, and southeast of the apex.

Interpretation of the regional differences in lithofacies indicates deposition by a distributary stream system whose apex was in south-central Utah and which spread sediments to the north, east, and southeast over a nearly flat plain. Irregularities on this plain near the Four Corners area and in west-central Colorado modified the distributary system, and therefore the wedge is not symmetrical.

Most uranium-vanadium ore deposits in the Salt Wash member occur in a lithofacies near the center of the wedge. This may be a genetic relation and can be explained as a function of transmissibility of the particular lithofacies. The ore deposits, however, are concentrated in a relatively small part of the central lithofacies. Because local geologic features such as structure or igneous intrusions might control the localization of ore deposits in the small area, the high degree of correlation of ore deposits and a certain lithofacies may be coincidental.

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Introduction

A lithofacies study of the Salt Wash member of the Upper Jurassic Morrison Formation was part of a detailed stratigraphic study of the Morrison Formation of the Colorado Plateau and adjoining regions made by the U. S. Geological Survey on behalf of the Division of Raw Materials of the U. S. Atomic Energy Commission. Tentative conclusions of the stratigraphic study were published by Craig, *et al.* (1955). They abstracted methods and conclusions reached by combined studies of lithofacies, regional stratigraphy, sedimentary structures, and sedimentary petrography. This paper is to illustrate methods and interpretations resulting from a particular lithofacies study rather than to present new conclusions about the Salt Wash member.

The term "lithofacies" in this paper denotes the total aspect of inorganic elements which furnish record of the depositional environment of a stratigraphic unit. This usage of "lithofacies" follows Krumbein (1948, p. 1923) and Kay (1947, p. 165) but differs from usage proposed by Moore (1949, p. 16); Moore would use the term "physiofacies" (p. 17) to denote the inorganic elements in a sedimentary rock. Moore would retain the term "lithofacies" to denote both inorganic and organic elements of a rock.

Specifically, "lithofacies", as used here, denotes the lithologic aspect as controlled by the thickness, the relative proportions, and the continuity of the two sedimentary types that constitute the Salt Wash member.

The purpose of the lithofacies study was to determine the regional variation in lithofacies of the Salt Wash member with the ultimate goals of determining: (1) some aspects of the depositional environment of the Salt Wash, and (2) relations between lithofacies and uranium-vanadium deposits in the Salt Wash.

Morrison Formation

Nomenclature

The Morrison Formation, defined originally by Cross (1894, p. 2), is present in much of the western interior of the United States. In the Colorado Plateau region the Morrison has been divided (in ascending order) into the Salt Wash member, defined by Lupton (1914, p. 127) and by Gilluly and Reeside (1928, p. 82), and the Recapture, Westwater Canyon, and Brushy Basin members, defined by Gregory (1938, p. 58–59). Table 1 shows a generalized section of the Morrison and adjacent formations in the Colorado Plateau region.

Salt Wash Member

The Salt Wash is best developed in southeastern Utah and in southwestern Colorado. As a recognizable unit the Salt Wash extends a short distance into northeastern Arizona and northwestern New Mexico where it disappears by depositional pinchout and by intertonguing and intergrading with the Bluff sandstone below and the Recapture member of the Morrison above. Southwest of a line trending northwestward through Lee's Ferry, Arizona, the Salt Wash was removed by erosion before the deposition of the Dakota sandstone. The western

limit of Salt Wash is concealed in most places by younger sedimentary rocks in the Wasatch Plateau but locally can be determined accurately along the southwest part of the San Rafael Swell. The northern and eastern limits of recognizable Salt Wash are near Vernal, Utah, and Glenwood Springs and Gunnison, Colorado. Salt Wash is a sandstone-bearing portion at the base of the Morrison Formation. Where this sandstone-bearing portion is absent, as in parts of northeastern Utah and northwestern and central Colorado, the Morrison Formation is not divided into members. Beds equivalent to Salt Wash, however, are probably present in the lower part of the undifferentiated Morrison Formation (Craig *et al.*, 1955). Figure 1 shows the outcrop pattern of the Morrison Formation in parts of Utah, Colorado, Arizona, and New Mexico and the areal extent of the Salt Wash member.

The Salt Wash member consists mainly of lenticular cross-laminated sandstone and conglomeratic sandstone interstratified with claystone, mudstone, siltstone, and structureless to horizontally laminated sandstone. Thin beds of limestone, locally containing fresh-water fossils, occur near the base of the Salt Wash in east-central Utah and west-central Colorado.

The cross-laminated sandstone is generally light-colored and ranges from fine- and medium-grained in Colorado to coarsely conglomeratic rock commonly containing lenses of cobble conglomerate in south-central Utah. The cross-laminated sandstone occurs as a single lensing bed 2–20 feet thick that extends less than 300 feet or as many cross-laminated sandstone beds which make up a composite bed. The composite beds are more abundant and generally much thicker than the single cross-laminated beds; some composite units form lenses that are more than 80 feet thick and extend several miles. All cross-laminated beds have gently undulatory to well-defined scour surfaces at the base.

The claystone, mudstone, siltstone, and structureless to horizontally laminated sandstone are mainly reddish brown. All gradations and mixtures of grain size between very fine-grained sand and clay are common among these rock types. Bedding in these rocks ranges from horizontal to gently lensing; fissility is rare.

The Salt Wash member is characterized by a steep ledgy outcrop. The cross-laminated

TABLE 1.—GENERALIZED SECTION OF MORRISON AND ADJACENT FORMATIONS IN THE COLORADO PLATEAU REGION

System	Formation	Member	Character
Cretaceous	Burro Canyon		Light-colored conglomeratic sandstone and green and red mudstone; mesa capping
Jurassic	Morrison	Brushy Basin	Variegated claystone and mudstone, some sandstone lenses; forms slopes
		Westwater Canyon	Light-colored sandstone with minor light-green mudstone; forms cliffs
		Recapture	Light-colored sandstone and red mudstone; generally less resistant than overlying and underlying members
		Salt Wash	Light-colored sandstone and red mudstone; forms cliffs and benches
	Bluff sandstone		Massive eolian sandstone; present only in Four Corners area
	Summerville		Evenly bedded red siltstone and very fine-grained sandstone
	Curtis		Glauconitic sandstone and greenish-gray shale

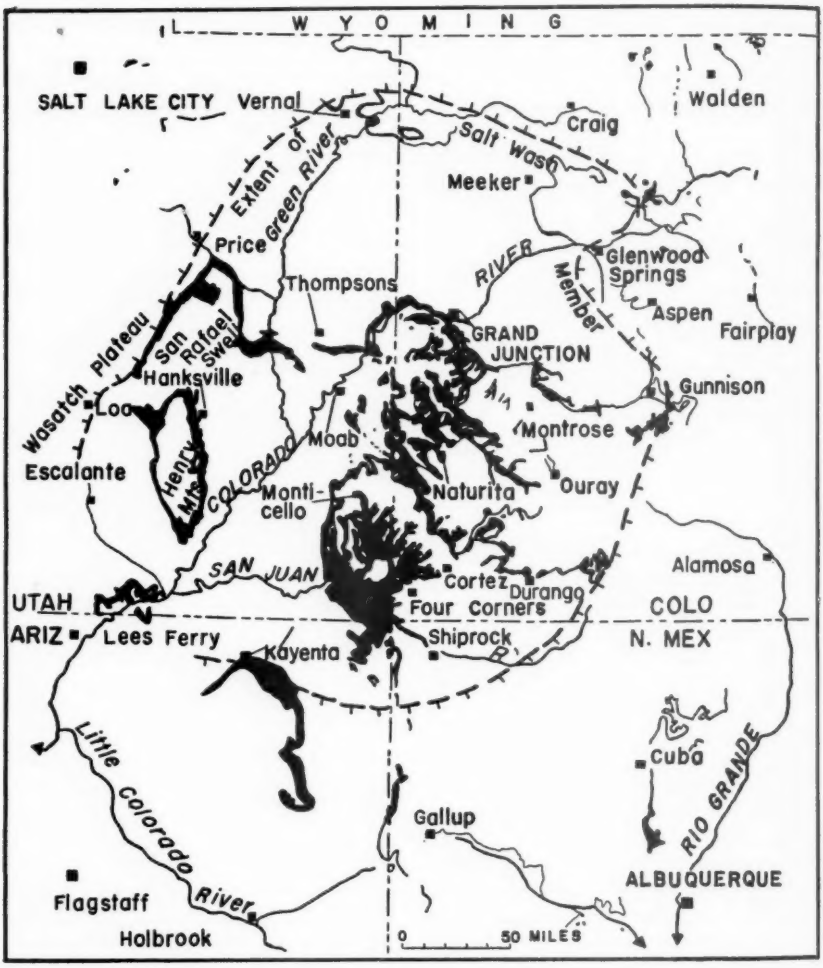


FIGURE 1.—MAP SHOWING OUTCROP PATTERN OF MORRISON FORMATION AND AREAL EXTENT OF SALT WASH MEMBER

sandstone forms steep-faced ledges; the finer-grained units form rubble-covered slopes between the ledges.

Because of the irregular assemblage of sandstone, siltstone, claystone, scour surfaces, and other sedimentary structures in the cross-laminated sandstone, the Salt Wash member is interpreted as a product of stream deposition (Craig *et al.*, 1955; Stokes, 1944; Mook, 1916).

Stratigraphic Relations of Salt Wash Member

The basal contact of the Salt Wash member in most areas is the base of the fluviatile deposits above the marine and marginal marine deposits of the Curtis and Summerville formations. The contact represents a change in depositional environment and is conformable in most places. In the San Rafael Swell and Henry Mountains area in Utah, however, contorted beds in the Summerville Formation are locally bevelled by the Salt Wash member; and over the crests of the salt anticlines in western Colorado the Salt Wash bevels older formations. Near Gunnison, Colorado, the Salt Wash rests on Precambrian rocks; and in Arizona, New Mexico, and in parts of Colorado and Utah near the Four Corners area the Salt Wash rests on the dominantly eolian Bluff sandstone. This second contact is a scour surface on the Bluff sandstone in most places, but locally fluviatile and eolian deposits intertongue.

In the Four Corners area most of the upper part of the Salt Wash member grades laterally into the Recapture member of the Morrison Formation. In other places the upper contact of the Salt Wash represents a change from a fluviatile environment to the dominantly lacustrine environment of the Brushy Basin member of the Morrison Formation. This change in environment is transitional; however, in many places where the basal deposits of the Brushy Basin are fluviatile a distinct composition and textural change between the Salt Wash and Brushy Basin members is noted.

Method of Study

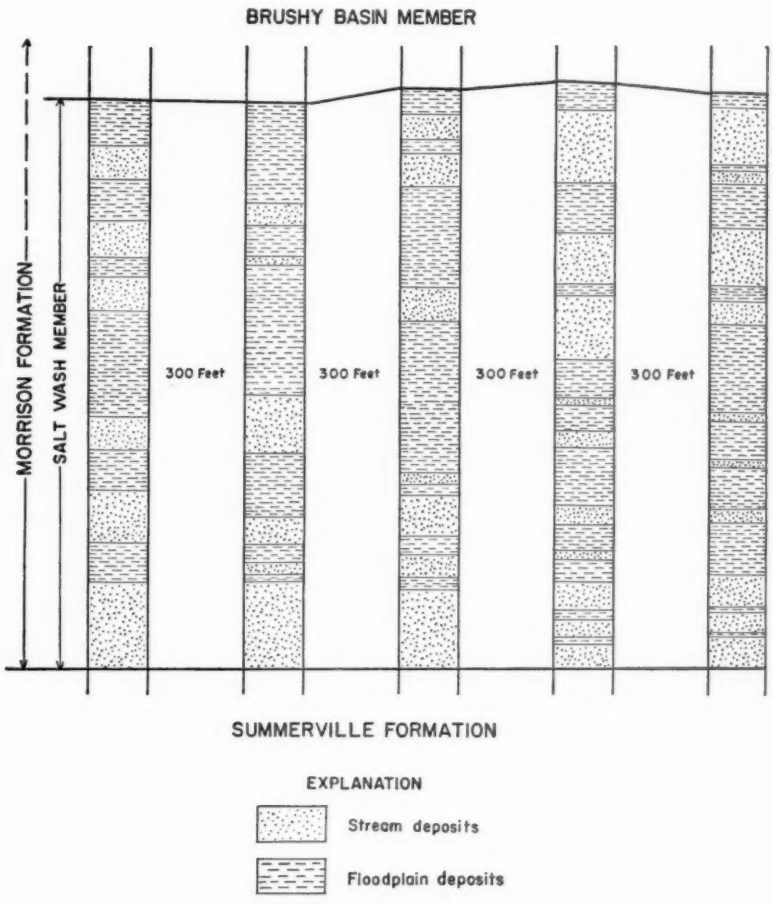
Classification of Sedimentary Units

Genetically the rocks in a fluviatile environment may be separated into stream deposits[1] and flood-plain deposits. The stream deposits consist of rocks which were deposited where sedimentation was noticeably influenced by water currents. In a fluviatile system these areas are restricted to channels and areas closely bordering the channels. The flood-plain deposits consist of rocks deposited from water in areas not noticeably influenced by current section. These areas are the relatively flat surfaces adjacent to stream channels where sedimentation from slack water occurs during and after floods.

The sandstone-shale classification of sedimentary units used in other lithofacies studies (Dapples *et al.*, 1948; Read and Wood, 1947; Sloss *et al.*, 1949) is not followed in this study for two reasons: (1) The genetic stream—flood-plain classification shows the relations between areas of current action and areas of no current action, whereas sandstone occurs in both stream and flood-plain deposits. Thus, drainage patterns based on stream deposits and flood-plain deposits should be more accurate than those based on sandstone and shale; (2) The genetic classification made possible a standardization of units measured. Gradation between claystone and sandstone is common, and sandy claystone or clayey sandstone is encountered in every measured section of the Salt Wash member. These units commonly form poorly exposed slopes between ledges of cross-laminated sandstone, and the relative amount of sandstone in the poorly exposed interval is difficult to determine. The problem of measuring the amount of sandstone in the clayey sandstone and sandy claystone was avoided by using a genetic classification, for these rock types are both flood-plain deposits.

The field distinction between stream and flood-plain deposits is clear in most cases. In this study a unit is considered a stream deposit if it is cross-laminated, has a basal scour surface, is composed of fine or larger grains, and is free of clay matrix. A unit is considered a flood-plain deposit if it is structureless or horizontally laminated and is composed of very fine-grained sand or smaller particles. Limestone, a minor rock type in the Salt Wash member,

[1] Many geologists would prefer "channel deposits" instead of "stream deposits" to describe sediments deposited where water currents influenced sedimentation. The writers use stream deposits to avoid conflict with a current Colorado Plateau use of "channel" to describe erosional features on the top of the Triassic Moenkopi Formation.



Vertical scale: 1 inch = 95 feet

FIGURE 2.—SECTIONS SHOWING LENTICULAR NATURE OF STREAM AND FLOOD-PLAIN DEPOSITS IN SALT WASH MEMBER OF MORRISON FORMATION

Sections measured at Head of Blue Canyon lithofacies locality, Mesa County, Colorado.

probably was deposited in shallow bodies of water on the flood plain and is included with the flood-plain deposits. The transitional rock in the stream—flood-plain deposits classification is very fine-grained well-sorted sandstone with poorly defined sedimentary structures. During field work most sandstone beds more than 2 feet thick were identified on the basis of the above criteria as stream deposits, and thinner sandstone beds were identified as flood-plain deposits. Consequently, very fine-grained sandstone beds 2 feet thick or less are considered flood-plain deposits and those more than 2 feet thick are considered stream deposits.

Collection of Data

Sampling technique.—Several sections were measured through the Salt Wash member at each lithofacies locality to calculate average thickness of the lenticular fluviatile deposits. Average thickness figures at each locality are necessary as a single section may not be representative; a section measured through the

fluviatile deposits may be considerably different in thickness of stream deposits and proportion of stream deposits to flood-plain deposits from one only 300 feet away (Fig. 2). Five sections continuity of the stream deposits were computed.

Sections were measured by Abney level and tape. The thickness and lithologic description

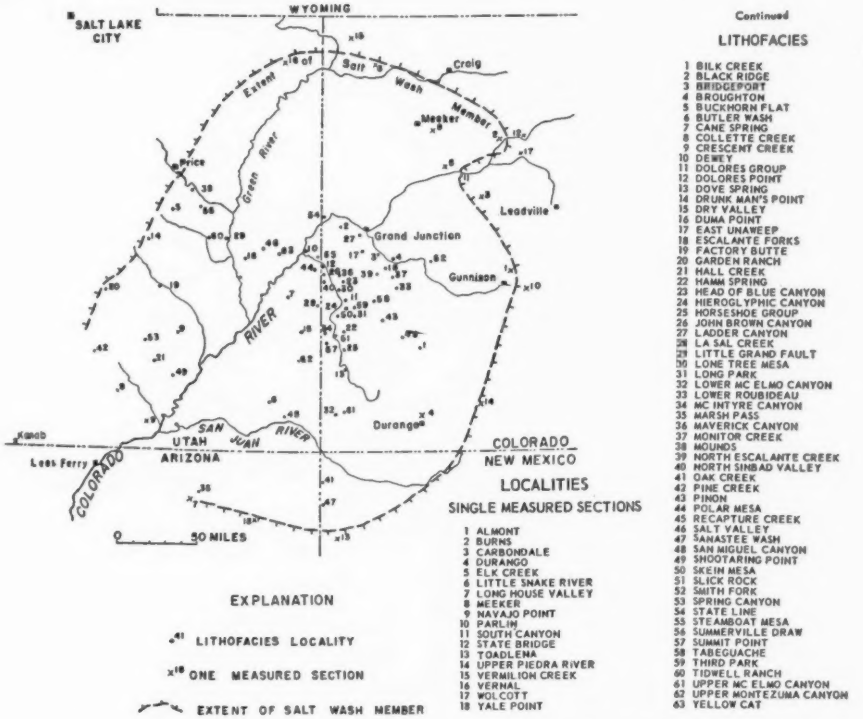


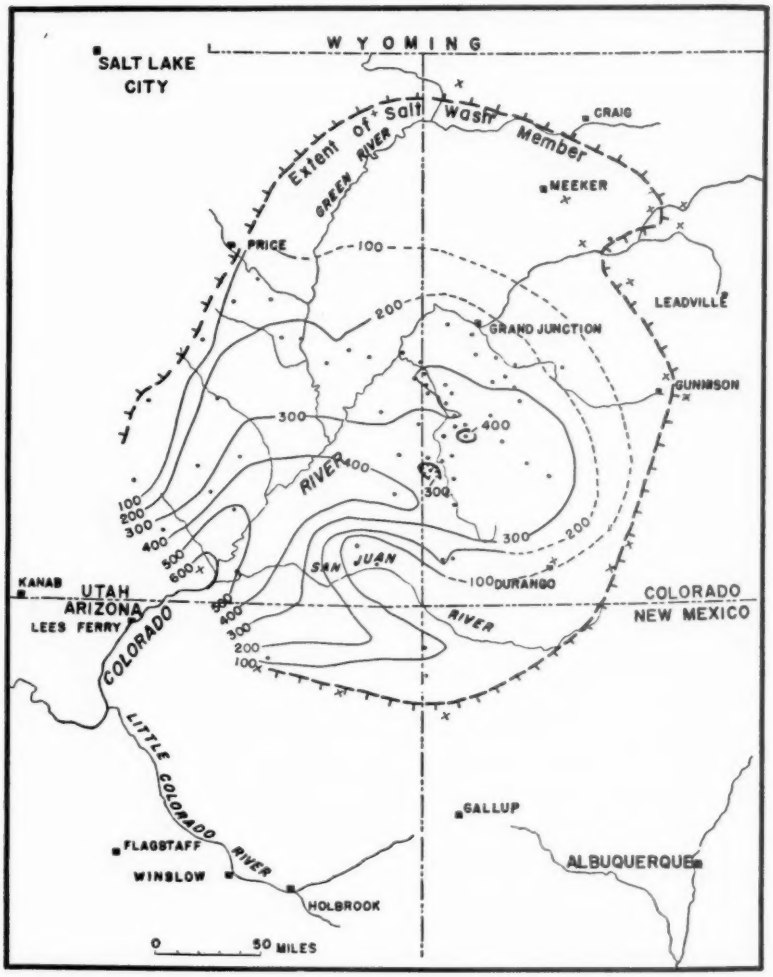
FIGURE 3.—MAP SHOWING AREAL EXTENT OF SALT WASH MEMBER OF MORRISON FORMATION

Includes location of lithofacies localities, and single measured sections.

evenly spaced over about 1200 feet of outcrop are thought to give a valid representation of the total lithologic aspect at each locality. Exposures and topography did not permit measuring five sections at all localities; however, all lithofacies localities represent at least three measured sections. The continuity of the sedimentary units at each locality can also be approximated if several sections are measured.

For each locality the average thickness of the Salt Wash member, the average thickness of contained stream deposits, the average thickness of the contained flood-plain deposits, the average percentage of contained stream deposits in the Salt Wash, and the relative of the units were recorded, and each unit was classified as either stream or flood-plain deposits as the section was measured.

Field work.—The area of study includes the southeastern quarter of Utah, the southwestern quarter of Colorado, and small parts of northwestern New Mexico and northeastern Arizona. All the Salt Wash member is within the area outlined, but no lithofacies measurements were made in northwestern Colorado and northeastern Utah where poor exposures of the Salt Wash prohibited lithofacies measurements by the sampling technique used. Field work started in western Colorado and was extended radially from this area with a decrease in concentration



EXPLANATION

- Lithofacies Locality
- One Measured Section
- 100 — Isopach Line, Dashed Where Inferred
- Extent of Salt Wash Member

Isopach Interval 100 Feet

FIGURE 4.—ISOPACH MAP OF SALT WASH MEMBER OF MORRISON FORMATION

of lithofacies localities away from western Colorado (Fig. 3).

Field work was begun in 1948 by L. C. Craig *et al.* who developed field techniques and made lithofacies measurements at 5 localities. In 1949, lithofacies measurements were made at 19 more localities by J. D. Ryan and the authors. The writers made lithofacies measurements at 39 additional localities in 1950 and 1951.

Data

Method of Presenting Data

Table 2 summarizes the data collected in the lithofacies study, and three isopach maps (Figs. 4, 5, 6) and two isolith maps (Figs. 7, 8) show the data graphically. In general, the method of converting the data to isopach and isolith maps follow methods of regional stratigraphic analysis proposed by Krumbein (1948, p. 1909–1923).

Except for use of single measured sections to delineate areal extent of the Salt Wash member, all data used in preparing the isopach and isolith maps were averages of measurements obtained from the several sections at each locality. The sections used to delineate the areal extent of the Salt Wash were measured by C. N. Holmes, L. C. Craig, and the writers.

Figures 4–8 represent conditions at the end of deposition of the Salt Wash member. The reconstruction involved drawing lines across areas where the Salt Wash either has been removed by erosion or is covered by younger rocks; however, the subjective element in drawing the isolith and isopach lines was lessened by plotting the lines at proportional intervals between control points. Isopach and isolith lines are not connected on the southwestern side of the maps, because in this area the Salt Wash was removed by Early Cretaceous erosion, and control points are not obtainable.

Isopach Map of Salt Wash Member

Figure 4 shows that the Salt Wash member is restricted to a nearly circular area. On the southwestern side a 140-mile gap in the circle indicates where the Salt Wash has been removed by erosion before deposition of the Dakota sandstone. The boundary of the circular area is represented by a hachured line which in part represents a true depositional pinchout and in part a grading of the Salt Wash into other units.

Isopach lines show that the Salt Wash member was deposited as a fan-shaped wedge. The apex of the fan, the area of greatest thickness, is in south-central Utah. Isopach lines, crudely concentric outward to the north, east, and southeast from the apex, are split into two distinct lobes by an area of thin Salt Wash in extreme southeastern Utah.

The southern lobe of the Salt Wash member is narrow and trends slightly south of east through northeastern Arizona and into New Mexico. The Salt Wash in this lobe thins uniformly toward its edge in New Mexico.

The northern lobe of the Salt Wash member is more extensive than the southern one. Isopach lines in the northern lobe form an asymmetric fan with an axis that curves eastward from the apex. Except in western Colorado, thinning in the northern lobe is rather uniform and is most rapid near the apex of the fan. Anomalous thicknesses, more than 400 feet at the Long Park locality and less than 300 feet at the Summit Point, Dolores Group, and North Sinbad Valley localities, prevent symmetry of the isopach lines in western Colorado. Isopach lines in the area of thin Salt Wash which separates the lobes of thicker Salt Wash form an oval whose long axis trends northwestward through the Four Corners area. In this area the Salt Wash overlies the thicker part of the Bluff sandstone which is predominantly wind-deposited. If the isopach lines were connected through this area of thin Salt Wash, they would form the shape of a symmetrical fan.

Isopach Map of Stream Deposits, Salt Wash Member

Figure 5 is similar to Figure 4 but shows more detail because of a smaller isopach interval. Greatest thickness of stream deposits is near the apex of the fan in south-central Utah. As in Figure 4, the isopach lines are divided into two lobes by an area of thin stream deposits in southeastern Utah. These lobes correspond in general to the lobes shown on Figure 4. The southern lobe is narrow and trends slightly

TABLE 2.—LITHOFACIES DATA FOR SALT WASH MEMBER OF MORRISON FORMATION

Lithofacies locality No.	Name	No. of sections at locality	Average thickness of stream deposits (Feet)	Average thickness of flood-plain deposits (Feet)	Average total thickness (Feet)	Average percentage of stream deposits	Percentage mean deviation of the thickness of stream deposits
1	Bilk Creek	3	140	197	337	42	5
2	Black Ridge	5	108	129	237	46	26
3	Bridgeport	4	78	167	245	32	9
4	Broughton	5	66	179	245	27	13
5	Buckhorn Flat	5	37	103	141	27	18
6	Butler Wash	5	42	27	68	64	9
7	Cane Spring	5	138	132	270	51	17
8	Collette Creek	3	240	129	369	67	6
9	Crescent Creek	4	263	110	372	71	5
10	Dewey	4	102	175	277	37	5
11	Dolores Group	5	136	152	288	47	15
12	Dolores Point	5	132	175	307	43	9
13	Dove Spring	3	204	158	361	56	6
14	Drunk Man's Point	5	15	68	83	18	10
15	Dry Valley	4	141	211	352	40	16
16	Duma Point	5	87	115	202	43	22
17	East Unaweep	4	88	194	282	31	15
18	Escalante Forks	3	116	181	296	39	19
19	Factory Butte	5	119	114	233	51	7
20	Garden Ranch	5	37	44	81	45	24
21	Hall Creek	4	249	86	335	74	7
22	Hamm Spring	4	147	198	341	44	11
23	Head of Blue Canyon	5	150	173	323	46	10
24	Hieroglyphic Canyon	4	183	154	338	54	5
25	Horsehoe Group	5	140	172	312	45	8
26	John Brown Canyon	5	117	199	316	37	11
27	Ladder Canyon	3	48	192	240	20	16
28	La Sal Creek	4	156	213	369	42	12
29	Little Grand Fault	4	103	156	259	40	13
30	Lone Tree Mesa	4	149	167	316	47	11
31	Long Park	5	278	134	422	68	5
32	Lower McElmo Canyon	5	140	127	267	52	21
33	Lower Roubideau	4	121	192	312	39	25
34	McIntyre Canyon	4	187	171	358	52	6
35	Marsh Pass	3	121	10	131	92	5
36	Maverick Canyon	4	134	192	326	41	17
37	Monitor Creek	3	135	164	299	45	6
38	Mounds	3	33	176	209	16	22
39	North Fork Escalante Creek	5	108	178	286	38	13
40	North Sinbad Valley	4	121	164	286	42	20
41	Oak Creek	5	170	27	196	87	6
42	Pine Creek	5	47	43	90	52	13
43	Pinon	4	193	128	322	60	8
44	Polar Mesa	5	104	189	293	36	24
45	Recapture Creek	4	32	0	32	93	6
46	Salt Valley	5	79	129	208	38	11
47	Sanastee Wash	3	14	37	51	28	29
48	San Miguel Canyon	5	116	134	316	47	16

TABLE 2.—*Continued*

Lithofacies locality No.	Name	No. of sections at locality	Average thickness of stream deposits (Feet)	Average thickness of flood-plain deposits (Feet)	Average total thickness (Feet)	Average percentage of stream deposits	Percentage mean deviation of the thickness of stream deposits
49	Shootaring Point	5	361	155	517	70	7
50	Skein Mesa	5	162	161	323	50	15
51	Slick Rock	5	165	161	329	50	16
52	Smith Fork	4	48	104	151	32	9
53	Spring Canyon	4	114	124	237	48	11
54	State Line	5	84	132	216	39	22
55	Steamboat Mesa	4	109	132	241	45	8
56	Summerville Draw	5	71	107	178	40	18
57	Summit Point	5	89	199	288	31	9
58	Tabeguache	4	148	154	302	49	3
59	Third Park	5	178	150	340	54	18
60	Tidwell Ranch	4	93	146	239	39	13
61	Upper McElmo Canyon	5	86	93	179	48	17
62	Upper Montezuma Canyon	5	200	220	420	49	15
63	Yellow Cat	3	97	153	250	39	7

south of east into New Mexico; the northern lobe is more extensive, and isopach lines show a crude asymmetric fan with an axis that curves eastward from the apex. In western Colorado isopach lines reveal two areas of anomalous thickness of stream deposits. Both areas are enclosed by isopach lines; however, the lines around the northwestern anomaly, the Long Park locality, indicate thickening of the stream deposits; and the line around the southwestern anomaly, the Summit Point locality, shows a thinning of stream deposits.

The area of thin stream deposits in southeastern Utah overlies the greatest thickness of the Bluff sandstone.

Isopach Map of Flood-Plain Deposits, Salt Wash Member

Figure 6 shows that the area of thickest flood-plain deposits is near the center of the area of Salt Wash deposition. In general, isopach lines are roughly concentric about the area of thickest flood-plain deposits, but the spreading of isopach lines from south-central Utah indicates a fanlike shape for the distribution of the flood-plain deposits.

The flood-plain deposits generally thin progressively from the center of the area, but two notable irregularities exist. One area, controlled by the Mounds and Little Grand Fault localities in east-central Utah, has anomalously thick flood-plain deposits, and another area, controlled by the Long Park, Pinon, and San Miguel Canyon localities in southwestern Colorado, has anomalously thin flood-plain deposits.

The 50-foot isopach line is not extended completely around the area of deposition because of the difficulty in distinguishing rocks of the Salt Wash member on the north and east sides of the Salt Wash area of deposition.

Isolith Map of Percentage of Stream Deposits in Salt Wash Member

Data for Figure 7 were prepared by computing the percentage of stream deposits in each section measured at a locality and then averaging the percentages of the several sections to obtain a value for the locality.

In general, the isolith lines form irregular concentric arcs outward from southeastern Utah. These arcs roughly parallel the boundary of the Salt Wash member except on the southern side where the isolith lines butt into the boundary.

The map shows that the highest percentage

of contained stream deposits is in southeastern Utah and northeastern Arizona. North of this area the percentage of contained stream deposits decreases rather gradually. To the northeast the percentage decreases irregularly and is not progressive, as several localities in western Colorado have relatively high percentages of stream deposits, and two, Summit Point and Dry Valley, have anomalously low percentages. South and southeast of the area of highest percentage of contained stream deposits the percentage decreases uniformly and relatively rapidly.

Isolith Map of Percentage Mean Deviation of Thickness of Stream Deposits in Salt Wash Member

The areal distribution of the relative continuity of the stream deposits in the Salt Wash member is shown in Figure 8. Quantitative data are computed in this manner: (1) the average of the thickness of stream deposits in all sections measured at the locality is determined, (2) the amount in feet that each section differs from the average of the locality is determined, (3) the differences are added (nonalgebraically) and divided by the number of sections measured at the locality, (4) the figure determined in (3) is divided by the average thickness of contained stream deposits at the locality, and (5) the figure determined in (4) is converted to a percentage by multiplying by 100.

The percentage mean deviation of thickness of contained stream deposits is a relative index of continuity. The relative index is desirable because it enables a comparison of continuity between localities which contain different thicknesses of stream deposits. In theory, a low-percentage deviation indicates high continuity between stream deposits, and a high percentage deviation indicates low continuity.

The index of continuity used in this study applies to the total Salt Wash member and is not an index for individual lenses of stream deposits in the Salt Wash. The index might possibly indicate high continuity where little or no continuity exists if sections of Salt Wash had the same thickness of stream deposits, but the stream deposits were not continuous between the sections. Possibilities that such conditions are reflected in the data are decreased by measuring several relatively closely spaced sections at each locality.

Figure 8 shows two areas of relatively high continuity of stream deposits. One area is in southeastern Utah and northeastern Arizona, the other is in western Colorado and east-central Utah. Both areas are enclosed by higher-value isolith lines that indicate a relatively low continuity of stream deposits away from the center of both areas. Isolith lines around the southern Utah and northeastern Arizona area of high continuity form roughly concentric arcs elongated northwestward. Continuity of stream deposits decreases progressively away from southeastern Utah in all directions except northeastward. Isolith lines around the Colorado and east-central Utah area of high continuity form a closed irregular lobate pattern.

Interpretation of Data

Method of Interpretation

To present the lithofacies of the Salt Wash on a single map requires a composite map of the three isopach and two isolith maps. Such a composite map is difficult to read and therefore none is presented. However, all interpretations are made on the basis of an integration of the five maps.

Depositional Environment

The first goal of this lithofacies study was to determine some aspects of the depositional environments of the Salt Wash member. The Salt Wash has long been considered a product of aggrading streams (Mook, 1916; Stokes, 1944), but little was known about the habit and distribution of those streams. The lithofacies data make possible some specific interpretations of the habit and distribution of the streams, and the interpretations indicate a specific depositional environment.

The lithofacies maps show the Salt Wash member to be a fan-shaped wedge of sedimentary rocks decreasing both in the relative proportion and continuity of stream deposits outward from the apex of the fan. Thus, the regional variations in lithofacies of the Salt

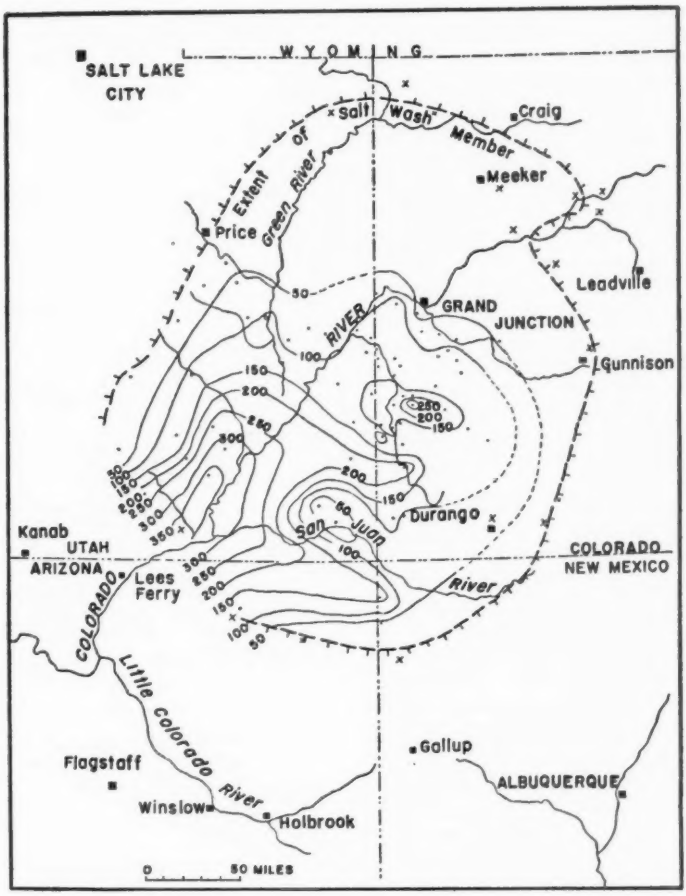
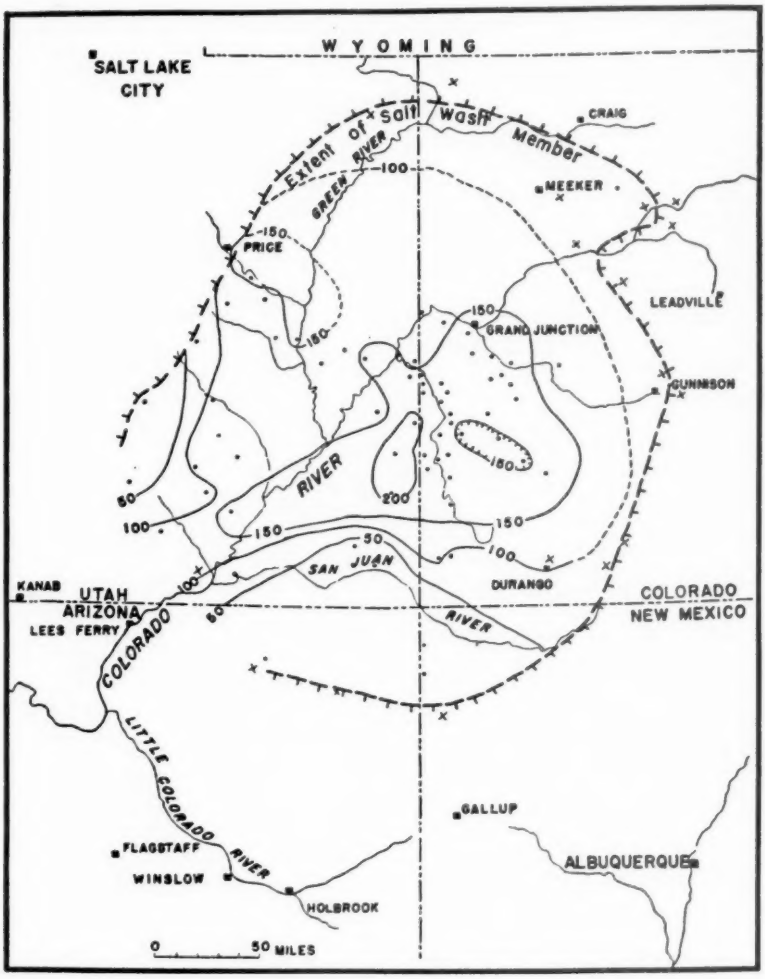


FIGURE 5.—ISOPACH MAP OF STREAM DEPOSITS IN SALT WASH MEMBER OF MORRISON FORMATION



EXPLANATION

• Lithofacies Locality

—50— Isopach Line, Dashed Where Inferred

× One Measured Section

Extent of Salt Wash Member

Isopach Interval 50 Feet

FIGURE 6.—ISOPACH MAP OF FLOOD-PLAIN DEPOSITS IN SALT WASH MEMBER OF MORRISON FORMATION

Wash can best be explained as resulting from a distributary drainage system which spread sediments north, east, and southeast from south-central Utah. The rough symmetry of the fan-shaped wedge of sedimentary rocks indicates that the distributary system was flowing over a nearly flat surface. In general, this surface was floored by marine and marginal marine deposits of the northward-retreating Late Jurassic sea. Therefore, the surface probably has a slight northward dip.

Field relations preclude the possibility that the distributary system existed as a delta in a marine environment; no intertonguing of Salt Wash rocks and marine or marginal marine rocks is known. On the other hand, sedimentary features and fossils support the terrestrial environment of the Salt Wash member.

Evidence is found in the Brushy Basin member of the Morrison Formation that, once established, the general pattern of the drainage system persisted throughout deposition of the Salt Wash member. The Brushy Basin is dominantly a lacustrine deposit, but, over much of the area of study, the basal deposits of the Brushy Basin are fluviatile. Although these deposits in the Brushy Basin are different in composition and texture from the underlying rocks of the Salt Wash, they apparently represent a continuation of the Salt Wash distributary system during early Brushy Basin deposition (Craig *et al.*, 1955, p. 157).

The maps compiled in the lithofacies study do not indicate an outlet for the water that transported and deposited the Salt Wash sediments. Possibly no outlet existed, and the water disappeared by seepage and evaporation, or possibly the Salt Wash streams drained by numerous outlets into the Late Jurassic sea north of the Salt Wash fan.

The symmetrical wedge of sedimentary rocks expected of a distributary system discharging on a flat surface is not shown by the lithofacies maps. Instead, the maps show irregularities that the writers believe can be correlated with irregularities on the surface of deposition.

Sand dunes that formed the Bluff sandstone made a highland in the Four Corners area. The highland did not persist throughout Salt Wash deposition, as the Salt Wash streams eventually built up their base level enough to flow across it. The highland, however, apparently split the Salt Wash distributary system into two lobes between which the Salt Wash member is thin because of restricted deposition.

Deviations from the pattern of a fan in west-central Colorado cannot be directly correlated with a particular geologic feature but probably reflect local warps in the Salt Wash basin of deposition; the lithofacies localities are too widely spaced to delimit the outlines of individual warps.

Salt anticlines in west-central Colorado and east-central Utah rose during Summerville deposition and continued rising until early in the time of Salt Wash deposition (Stokes and Phoenix, 1948). Upward movement of the salt anticlines was expressed topographically in the Salt Wash plain of deposition. The low-gradient Salt Wash streams probably were dammed in many places and probably deposited much of their load near the flanks of the salt anticlines.

In western Colorado other deviations from the fan-shaped pattern formed by distributary streams may be related to slight topographic relief of the Uncompahgre highland. During the time of Salt Wash deposition Precambrian rocks of the Uncompahgre highland extended from north-central New Mexico through Gunnison, Colorado. Just north of Grand Junction, Colorado, the highland passed into a structural terrace and had no topographic expression during the time of Salt Wash deposition (Holmes, 1951). The Uncompahgre highland did not stand high enough to contribute more than a minor amount of sediment to the Salt Wash, but it possibly formed a barrier to the low-gradient eastward-flowing Salt Wash streams. Such a barrier would cause unequal distribution of sediments in the Salt Wash fan as a more than normal amount of sediment would be deposited on the west side and a less than normal amount on the east side. The thicker Salt Wash, which had a relatively high proportion and continuity of stream deposits centering at the Long Park locality, may reflect the effect of the Uncompahgre highland.

Another possible effect of the Uncompahgre highland on the Salt Wash wedge is shown by the sharp curve in the boundary of the Salt

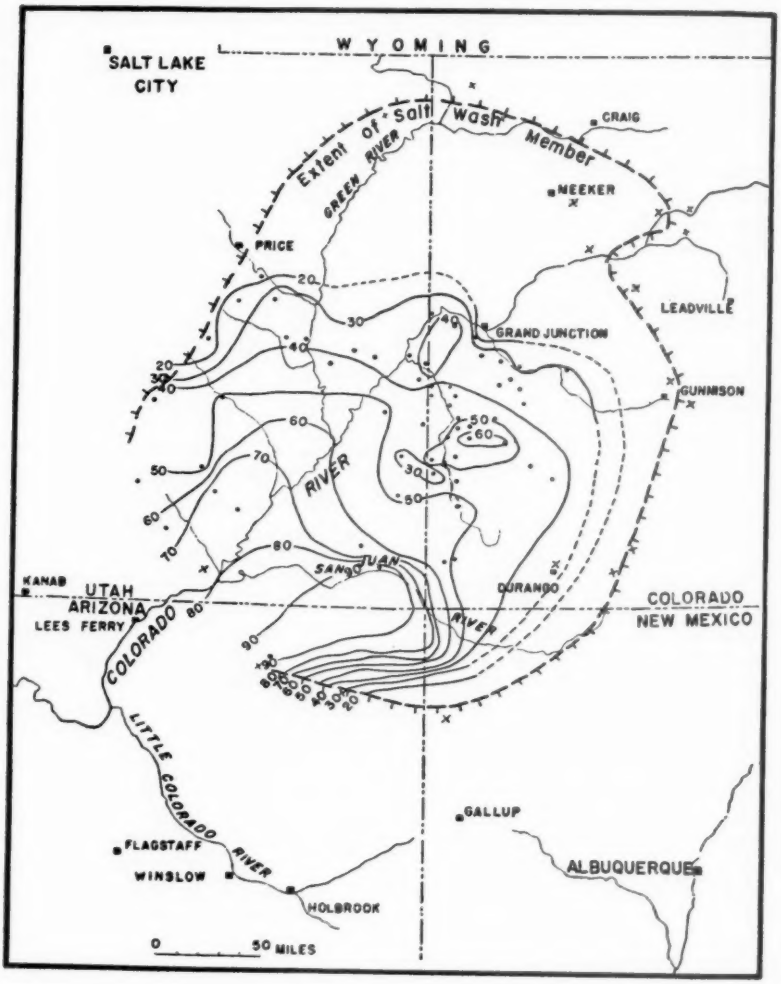


FIGURE 7.—ISOLITH MAP OF PERCENTAGE STREAM DEPOSITS IN SALT WASH MEMBER OF MORRISON FORMATION

Wash member between Leadville and Meeker. Sedimentary rocks of the Salt Wash member are present at the Burns and Elk Creek sections, but not at the South Canyon, Wolcott, or State Bridge sections. Southward from the South Canyon section there is no recognizable Salt Wash until the Almont section. This curve in the margin of the Salt Wash member does not correspond to the symmetrical margin of a fan as the northeastern margin extends farther from the apex than the eastern margin. The asymmetry may have been caused by the configuration of the Uncompahgre highland. Salt Wash streams northwest of Grand Junction, Colorado, would not have been affected by the highland barrier. Streams southeast of Grand Junction would decrease in gradient and lose part of their capacity to transport sediments as they neared the highland. Therefore, other things being equal, streams northwest of Grand Junction could transport sediments farther from the apex of the distributary system than streams southeast of Grand Junction. This arrangement could produce an asymmetrical curve at the margin of the Salt Wash fan.

The interpretation of the depositional environment of the Salt Wash member presented here emphasizes a single source area of sediments and a single distributary drainage system to spread the sediments. This interpretation is based on the maps described in this report. The writers are pleased that other studies on the Salt Wash support their interpretations.

Conglomerate dominates the Salt Wash at the apex of the fan and decreases away from the apex (Craig *et al.*, 1955, p. 138). A study of cross-laminations in the stream deposits in the Salt Wash member indicates that Salt Wash streams formed a distributary pattern whose apex was near the apex determined by lithofacies study (Craig *et al.*, 1955, p. 145). Some aspects of the Salt Wash deposition as determined by other stratigraphic studies, however, are not reflected by the lithofacies maps. Conglomerate in the Salt Wash member near the Garden Ranch lithofacies locality and petrographic studies of Salt Wash rocks indicate that the Salt Wash probably had minor contributions of sediment from west-central Utah (Craig *et al.*, 1955, p. 147–150). These minor source areas are not reflected in the lithofacies maps.

Relations of Lithofacies to Uranium-Vanadium Ore in Salt Wash Member

The second goal of the lithofacies study was to determine any relation between lithofacies of the Salt Wash member and the uranium-vanadium deposits in the Salt Wash.

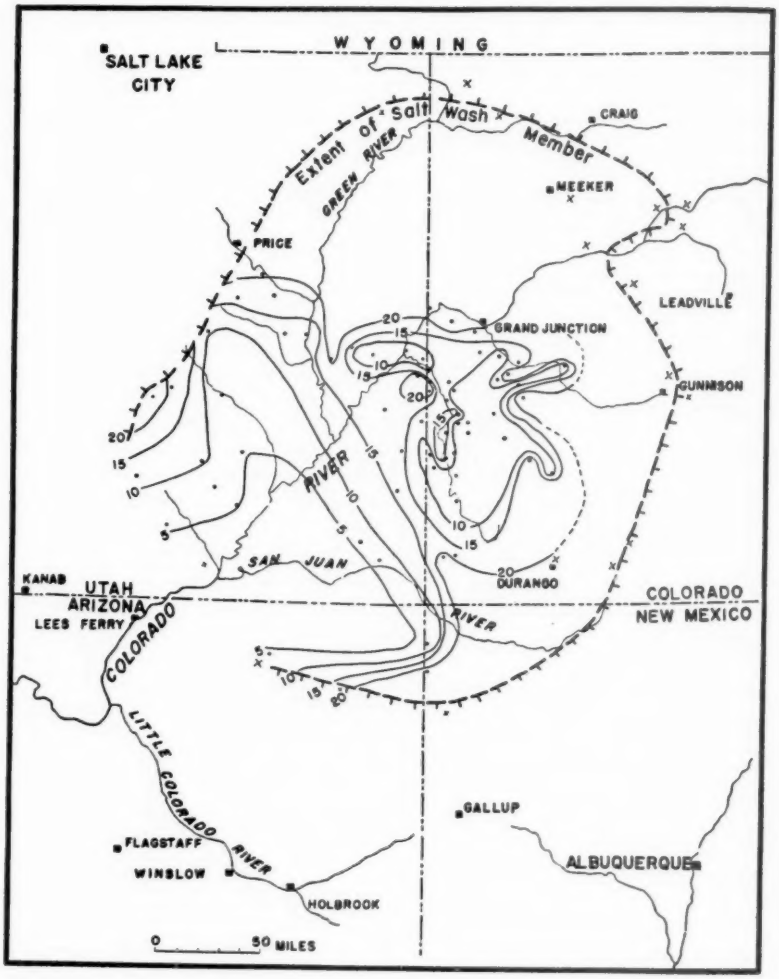
Uranium and vanadium ore is mined from the Salt Wash member in many places in Colorado, Utah, and Arizona. The ore occurs as irregular tabular masses in the thicker part of stream-deposited sandstone lenses and, in general, is restricted to a single zone within the Salt Wash member. The ore deposits have been described in detail by Fischer (1942).

If the uranium and vanadium were transported in solution through the Salt Wash member, the movement of the metal-bearing solution would be affected by factors controlling the transmissibility of the Salt Wash. The stream deposits are more permeable than the flood-plain deposits; therefore, major factors controlling the transmissibility of the Salt Wash are the thickness, continuity, and relative proportion of stream deposits in the Salt Wash. Favorable conditions of these three factors might form traps for concentrations of metal-bearing solutions by density stratification or related phenomena.

Three conditions of transmissibility in the Salt Wash member and the possible relations of transmissibility to localization of ore deposits are illustrated in Figure 9. Permeability of the stream deposits and hydraulic gradient are assumed to be the same in each case illustrated.

A low proportion of stream deposits to flood-plain deposits and little continuity of the stream deposits is shown by *A* in Figure 9. Movement of solutions through the stream deposits is impeded because little continuity exists between individual lenses of stream deposits. The opportunity for concentrations of dissolved metals from weak solutions is unlikely because little solution passes through the stream deposits.

The assumed ideal lithofacies for concentration of ore minerals in the stream deposits of the Salt Wash member is shown by *B* in Figure 9. Stream deposits, which constitute



EXPLANATION

- Lithofacies Locality
- Isolith Line, Dashed Where Inferred
- One Measured Section
- Extent of Salt Wash Member

Isolith Interval 5 Percent

FIGURE 8.—ISOLITH MAP OF PERCENTAGE MEAN DEVIATION IN THICKNESS OF STREAM DEPOSITS IN SALT WASH MEMBER OF MORRISON FORMATION

about half the thickness, are continuous but slightly lenticular. Movement of metal-bearing solutions would be restricted to the stream deposits and the movement of the solution is locally impeded by lenticularity of the stream deposits. Dissolved ore metals might be concentrated at favorable places controlled in detail by sedimentary structures, because the stream deposits are continuous enough to allow inflow of new solution.

Stream deposits that have high continuity and constitute about three-fourths of the thickness are shown by *C* in Figure 9. This lithofacies may be unfavorable for the accumulation of ore deposits. Movement of metal-bearing solutions through the Salt Wash member would not be restricted nor would movement of the solutions be restricted by irregularities in the stream deposits. The metal-bearing solutions could pass freely through this lithofacies, and there would be little chance for concentration of the ore metals.

Regionally, transmissibility of the Salt Wash member probably is greatest near the apex of the fan in south-central Utah. The stream deposits show greatest thickness and continuity and constitute over half the thickness of the Salt Wash in this area. Transmissibility of the Salt Wash probably decreases away from the apex as do the thickness, continuity, and relative proportion of stream deposits. Near the margin of the fan the transmissibility of the Salt Wash is probably lowest, as stream deposits constitute little of the total thickness, and little continuity exists between individual lenses. The assumed ideal conditions of transmissibility for accumulation of ore deposits are in the central part of the Salt Wash wedge.

Quantitative relation of uranium-vanadium ore to lithofacies of the Salt Wash member was obtained by plotting lithofacies parameters of ore-bearing localities in relation to all quantitative lithofacies data. For this purpose a lithofacies locality was considered an ore-bearing locality if at least 100 tons of rock containing 0.1 per cent of uranium oxide was or is known to be within 1 mile of the locality. Of the 63 lithofacies localities, 22 are ore-bearing localities.

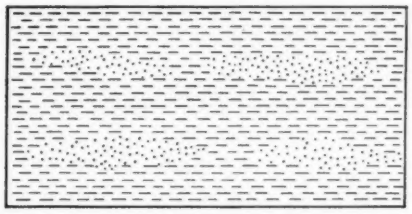
Of the 22 ore-bearing lithofacies localities 18 occur where the Salt Wash member is more than 240 feet thick and composed of 40–55 per cent stream deposits, contains 90–200 feet of stream deposits, and has a 5–18 per cent mean deviation in the thickness of stream deposits. As 18 of 37 localities included in the lithofacies outlined by the parameters above are ore bearing and only 4 of 27 localities not in this lithofacies are ore bearing, the parameters seemingly delimit a lithofacies that is more favorable for the accumulation of ore deposits than other lithofacies. The favorable lithofacies for ore deposits is a statistical concept. If parameters from all 22 ore-bearing localities are used to determine a favorable lithofacies for ore deposits, the resulting lithofacies includes about 90 per cent of the Salt Wash wedge.

Figure 10 shows the areal distribution of sedimentary rocks of the Salt Wash member that conform to the quantitative lithofacies limits outlined as most favorable for ore deposits. The area outlined is, in general, the central part of the Salt Wash wedge. The figure also shows that most major ore deposits in the Salt Wash (Fischer, 1942, Pl. 1) conform to the limits of favorability as determined by the lithofacies parameters. This may be a genetic relationship explained as a function of the transmissibility outlined in *B* of Figure 9 if a uniform source of uranium and vanadium is assumed.

Although Figure 10 shows an excellent correlation between a certain lithofacies and ore deposits, the correlation may be coincidental instead of genetic.

In the assumed ideal lithofacies for accumulation of ore deposits, the ore deposits should not be confined to a particular stratigraphic zone. Actually, ore deposits in the Salt Wash member in a given mining district are generally confined to one stratigraphic zone of stream deposits. Therefore, an exact correlation between total lithologic aspect of the Salt Wash and ore deposits confined to a single zone in the Salt Wash may be questioned. The strongest possible correlation is that the favorable lithofacies may represent a depositional environment in which single zones favorable for deposition of ore are more likely to occur than in other lithofacies.

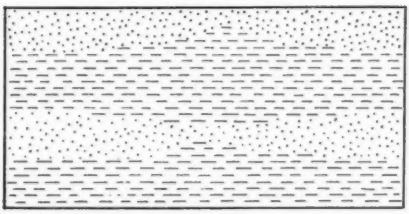
The weakest possible correlation is that there is no genetic relationship between lithofacies



Noncontinuous stream deposits. Movement of solutions is hindered

Theoretically unfavorable for accumulation of ore minerals

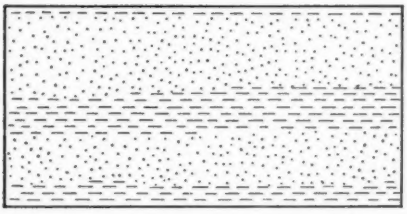
A



Continuous stream deposits with irregularities that restrain solutions in places

Theoretically favorable for accumulation of ore minerals

B



Continuous stream deposits with no irregularities to interrupt the movement of solutions

Theoretically unfavorable for accumulation of ore minerals

C

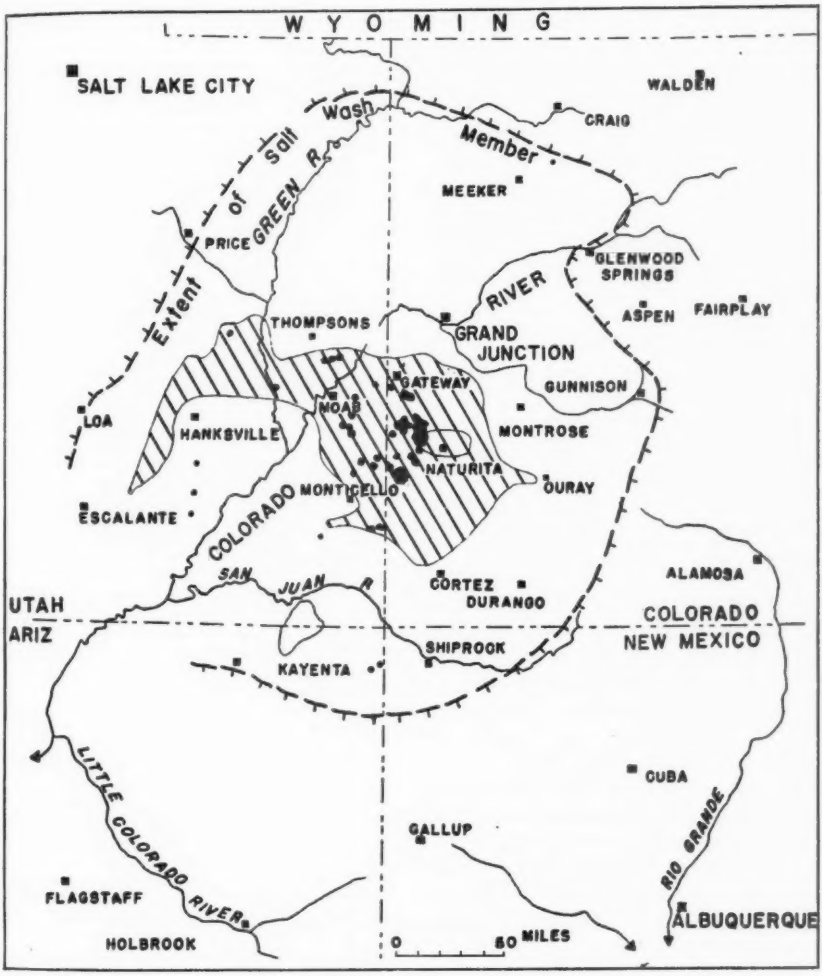
EXPLANATION

Stream deposits

Flood plain deposits

FIGURE 9.—DIAGRAM SHOWING CONDITIONS OF TRANSMISSIBILITY IN SALT WASH MEMBER OF MORRISON FORMATION

and location of ore deposits. Figure 10 shows that ore deposits are not distributed uniformly through the favorable lithofacies. Instead, most ore deposits are concentrated in east-central Utah and west-central Colorado, an area which forms a small part of the favorable lithofacies; and some major ore deposits occur in lithofacies here considered unfavorable for ore deposits. Possibly the ore deposits are related to local structural features, localized hydrothermal actvity, or other unknown local features. Therefore, the good correlation



EXPLANATION

• LOCATION OF MAJOR URANIUM-VANADIUM DEPOSIT

Area of favorable Salt Wash as determined by Lithofacies study (Salt Wash Member over 240 feet, stream deposits comprise 40 to 55 percent of member, thickness of stream deposits is 90 to 200 feet, percentage mean deviation of stream deposits is 5 to 18 percent)

Extent of Salt Wash Member

FIGURE 10.—MAP SHOWING THE LOCATION OF MAJOR URANIUM-VANADIUM DEPOSITS In Salt Wash member of Morrison Formation and areas of favorable Salt Wash as determined by lithofacies study.

between ore deposits and a particular lithologic aspect of the Salt Wash member may be coincidental.

Summary

Regional variations in lithofacies (lithologic aspect) of the Salt Wash member are the basis for interpretations of the depositional environment of the Salt Wash member. The regional variations in lithofacies indicate deposition by a distributary stream system from an apex in south-central Utah. The distributary system spread sediments to the north, east, and southeast over a nearly flat plain forming a fan-shaped deposit. The symmetry of the Salt Wash wedge of sedimentary rocks was modified by irregularities in the surface of deposition in the Four Corners area and in west-central Colorado.

Theoretical considerations of the lithofacies of the Salt Wash member indicate that a lithofacies near the center of the wedge should be ideal for the accumulation of the uranium-vanadium ore deposits. Quantitative parameters of Salt Wash lithofacies indicate that many uranium-vanadium ore deposits are restricted to this central lithofacies. The ore deposits are not distributed uniformly through this central lithofacies, however; instead most known ore deposits are concentrated in a small part of the central lithofacies zone. Because local geologic features not related to Salt Wash depositional environment might control the distribution of the ore deposits, the high degree of correlation of the central lithofacies and ore deposits may be coincidental.

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U. S. Geological Survey, Grand Junction, Colorado

Manuscript Received by the Secretary of the Society, November 28, 1955

Publication Authorized by the Director, U. S. Geological Survey